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SEVERE THUNDERSTORM FORECASTING --

AN OPERATIONAL REVIEW

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Preface

This publication represents the first of a two-stage project aimed at the establishment of a comprehensive NWS field guide to forecasting severe thunderstorm occurrence. In its present form, the publication serves as a brief survey of pertinent recent literature regarding severe thunderstorm structure, detection and forecasting. It is organized so as to simulate the methodology or approach an average forecaster might take in preparing for and actually working a severe thunderstorm event. The main group targeted by this publication and the planned final version is the new forecaster or Intern whose knowledge of severe thunderstorm structure and forecasting may be quite limited. However, it is recognized that recent advances in the science can take their toll on experienced forecasters as well. Thus, we hope that this guide also might serve to maintain proficiency for these personnel.

The final version of the publication will incorporate a more effective training approach. It will be modular in design, making it easier to digest information and use in an operational environment. Each chapter will include a list of training objectives along with a suggested approach to studying the information. It is possible that a glossary of severe thunderstorm terms also might be included by chapter. Finally, being basically a field guide written by operational forecasters, we hope that we might provide some insight regarding operational demands of the forecast office environment that, to date, have received limited attention.

We recognize that there are a number of forecasters in the field that have considerable expertise in working severe thunderstorm events. As such, we certainly welcome any constructive comments or suggestions they might be willing to offer in the preparation of the final guide. Please address all correspondence to:

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Your help will be much appreciated.

Contents

Chapter 1

SEVERE THUNDERSTORM CLIMATOLOGY

The first step in successfully forecasting the onset of any weather phenomena is to have some idea, in a statistical sense, of when atmospheric conditions are favorable. If for no other reason, it heightens the awareness of the forecaster and gives him or her some sense of what to look for. This is especially critical in the forecasting of severe thunderstorms and their damaging offspring. In this section, we will look at the climatology of severe thunderstorm occurrence; specifically which types of storms are most likely to occur, the sections of the country most vulnerable to their occurrence, and when they are most likely to develop. We will finish this section by making some climatological comparisons among the by-products of the severe thunderstorm, i.e. damaging wind, hail and tornadoes.

Classification of Severe Thunderstorms

A great deal of work has been done in recent years in uncovering the structural characteristics of severe thunderstorms. The following section on "Severe Thunderstorm Models" will discuss many of these findings in some detail. But, for now let it suffice to say that there are basically three types or structures of a severe thunderstorm: the single cell pulse storm, the multicell storm, and the supercell storm. There are certain subcategories inherent in each type, especially the multicell storms, but we leave those details for the following section.

Naturally, each type of severe thunderstorm starts out as a single cell non-severe storm and evolves at various rates depending upon a number of factors into a severe storm. From a climatological viewpoint, it is only important to note the relative frequency at which this occurs. This is outlined in Figure 1-1. Admittedly, this is a rather complex diagram, but careful examination reveals some interesting information. First, in the upper portion of the diagram, notice that the majority of thunderstorms occur as either single cell or multicell. Less than 50 percent of these evolve into severe storms, with only a very small percentage actually becoming supercells. As shown in the lower part of the diagram though, we must not be fooled by the small frequency of occurrence of supercells, since they produce a sizable percentage of the damage from severe thunderstorms. There is a dramatic point to be made here for the forecaster. It is time well-spent to learn the clues to supercell development. Even though they are very infrequent, you can bet that they will produce considerable damage once they do develop.

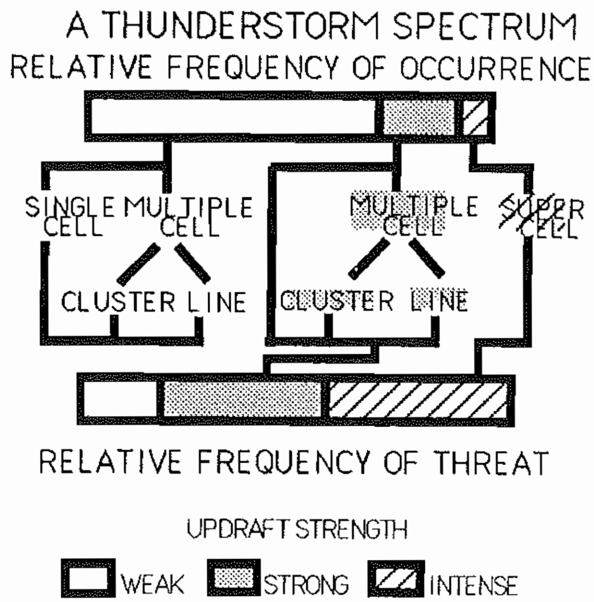


Figure 1-1. Relative frequency of thunderstorm occurrence and threat categorized by updraft strength.

Spatial Characteristics of Severe Thunderstorms

The development of severe thunderstorms is an evolutionary process initiated from a small, non-severe cell. As such, one would expect that the spatial distribution of severe storms and the weather they produce would be a subset of total thunderstorm distribution. With this in mind, it would seem unnecessary to rehash information that is printed in virtually every meteorological textbook addressing the subject of thunderstorms. Let's do however, mention that South Texas does experience a number of severe thunderstorm episodes, mostly concentrated in the spring months of March, April and May. The coastal plains of Southeast Texas experience the highest percentage of these occurrences where moisture off the Gulf of Mexico most efficiently fuels their development.

Severe thunderstorms and their associated damaging offspring develop into either isolated or multiple events. Isolated events generally are short-lived. Multiple events, on the other hand, have longer durations and are responsible for more deaths. Studies of U.S. severe weather deaths from 1967 through 1973 indicated that 87% of those deaths occurred during multiple events.

Multiple events appear clearly to exhibit two separate spatial patterns, either forming along corridors or in clusters (Moller, 1980). Corridor events take the form of elongated, elliptically-shaped areas. Events of this type tend to develop in the more moisture-rich parts of the country. Their peak occurrence is in April and May and is associated with stronger upper-level dynamics where the 500-mb wind averages 45 to 50 kts.

Cluster events are characterized by smaller, circular-shaped areas and the tendency to develop over drier portions of the country, such as the Plateau region of the west. Events of this type seem to depend heavily on low-level thermodynamics, and as such, are associated with weaker upper-level dynamics where the 500-mb wind averages 25 to 30 kts. Their peak occurrence is a bit later, in May and June.

Temporal Characteristics of Severe Thunderstorms

From an annual viewpoint, the occurrence of severe thunderstorms is closely tied to insolation, the movement of the polar jet stream, and associated synoptic-scale disturbances. For these reasons, severe weather is often seen to erupt first in late winter over the Southeast U.S. and the Gulf Coast where warm, moist air is most readily available to fuel development. By spring, severe weather occurrences spread westward into the Central Plains and southwest portions of the country. Finally, as we move into summer, northcentral sections of the U.S., including the Great Lakes, become more susceptible to severe thunderstorm occurrences.

Diurnally, severe thunderstorms can occur at any time of the day or night. This is especially true if the major lifting mechanism to trigger development results from synoptic-scale forcing (e.g. surface fronts, strong upper disturbances, etc.). However, statistics indicate that the time of greatest severe thunderstorm threat is in the late afternoon and early evening hours, or between 1600 and 1800 local time. This obviously supports the importance of the role of low-level heating in storm development. More will be said about this in subsequent sections.

Associated Phenomena

By definition, severe thunderstorms are known to produce damaging wind of 50 knots or greater, large hail of 3/4 inch or larger, and/or tornadoes. While all of these by-products can and often do occur near-simultaneously within a given thunderstorm or cluster of thunderstorms, each phenomenon does exhibit a distinct climatological frequency and time of occurrence. For example, as shown in Figure 1-2, severe thunderstorms that produce wind damage other than tornadoes occur primarily in the summer months of June and July (Doswell et al., 1983). Hailstorms, on the other hand, reach a peak in May and June. Diurnal differences (Figure 1-3) are not as apparent, with both phenomena reaching peaks near sunset.

Spatial and temporal variations of tornado occurrence closely resemble that of severe thunderstorm occurrence. In other words, tornadoes are most frequent in late winter over the Southeast U.S. and the adjacent Gulf Coast. Spring months find the tornado maximum over central and southwest portions of the country, with it shifting into northcentral sections by summer.

The hours of peak tornado occurrence, once again, are between the hours of 1600 and 1800 local time. However, a secondary peak is apparent in the early morning hours, near the time of sunrise. Frequencies, though, are far less than for the afternoon peak.

The direction of movement of tornadoes is usually in accordance with the average mid- and upper-level wind flow, i.e. to the northeast with an average forward speed of 25 mph. However, they can make almost instantaneous changes in movement; slowing down, becoming stationary or possibly accelerating to speeds upward of 70 mph. All the while, they can be looping or turning in any combination of ways.

Tornadoes, just like their parent severe thunderstorms, develop over a spectrum of sizes and intensities. Scales have been developed for the sake of categorizing these characteristics. Dr. Theodore T. Fujita devised a scale of windspeeds (F0 through F5) and Dr. Allen Pearson has developed a scale of path widths and lengths. Pictorial examples are given in Figure 1-4. In general, tornadoes can be lumped into three very broad classes.

F0 and F1 tornadoes are "weak", with path lengths usually less than one mile and widths less than 150 yards. They are, by far, the most common of tornadoes, making up about 70% of all tornadoes. Their duration generally is less than 2 minutes. This, in addition to their small size, make detection both by radar and trained spotters very difficult if not impossible. As a result, our warning capability is quite poor. However, their maximum wind is in the range of 70 to 100 mph, and consequently can

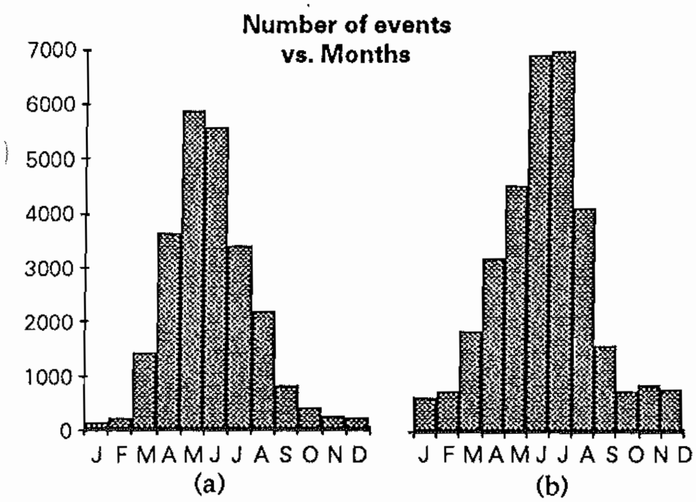


Figure 1-2. Annual distribution of severe thunderstorms that produce (a) wind damage, and (b) large hail.

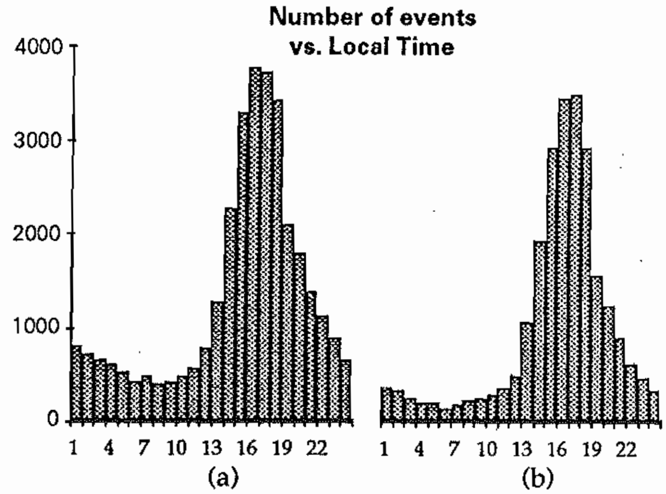


Figure 1-3. Diurnal distribution of severe thunderstorms that produce (a) wind damage, and (b) large hail.

produce considerable damage. Fortunately, very few deaths have been attributed to this class of tornadoes.

F2 and F3 tornadoes are referred to as "strong". They have path lengths on the order of 1 to 10 miles and widths ranging from 150 to 500 yards. Their average lifespan is from 3 to 20 minutes, and they can cause tremendous damage with wind speeds of 100 to 200 mph. Strong tornadoes account for almost 30% of all reported tornado occurrences and almost 30% of related deaths. Due to the larger size and longer lifespan of F2 and F3 tornadoes, both radar and spotters have had some success in detecting and tracking this class of tornadoes. As such, our warnings appear more timely.

Finally, there are the F4 and F5, or "violent" tornadoes. They make up less than 2% of all tornado occurrences, but account for approximately 68% of tornado-related deaths. Path lengths can exceed 100 miles in extremely rare cases, as evidenced by the Tri-State twister of 1925. This tornado traveled across Missouri and Illinois before dissipating in Indiana. The death count from this devastating storm was an astounding 800 people. Violent tornadoes can be immense in size with path widths up to one and a half miles (e.g. the Wichita Falls, Texas storm in 1979). Their average duration can be anywhere from a few minutes to several hours. Wind speeds can reach extremes of 300 mph when taking into account a 250 mph rotational speed coupled with a possible 50 mph translational speed from the parent thunderstorm. The highest tornadic wind was 221 mph (measured by photogrammetry) in the Xenia, Ohio tornado of 1974. Fortunately, radar and spotters have experienced a great deal of success in detecting and tracking these storms, and warnings are usually issued well in advance.

(F0) Gale tornado (40-72 mph): Light damage
Some damage to chimneys; break branches off trees; push over shallow-rooted trees; damage sign boards.

(F1) Moderate tornado (73-112 mph): Moderate damage
The lower limit (73 mph) is the beginning of hurricane wind speed; peel surface off roofs; mobile homes pushed off foundations or overturned; moving autos pushed off the roads.

(F2) Significant tornado (113-157 mph): Considerable damage
Roofs torn off frame houses; mobile homes demolished; boxcars pushed over; large trees snapped or uprooted; light-object missiles generated.

(F3) Severe tornado (158-206 mph): Severe damage
Roofs and some walls torn off well-constructed houses; trains overturned; most trees in forest uprooted; heavy cars lifted off ground and thrown.

(F4) Devastating tornado (207-260 mph): Devastating damage
Well-constructed houses leveled; structure with weak foundation blown off some distance; cars thrown and large missiles generated.

(F5) Incredible tornado (261-318 mph): Incredible damage
Strong frame houses lifted off foundations and carried considerable distance to disintegrate; automobile-sized missiles fly through the air in excess of 100 m; trees debarked; incredible phenomena will occur.

(F6-F12) (319 mph to Mach 1, the speed of sound): The maximum wind speeds of tornadoes are not expected to reach the F6 wind speeds.

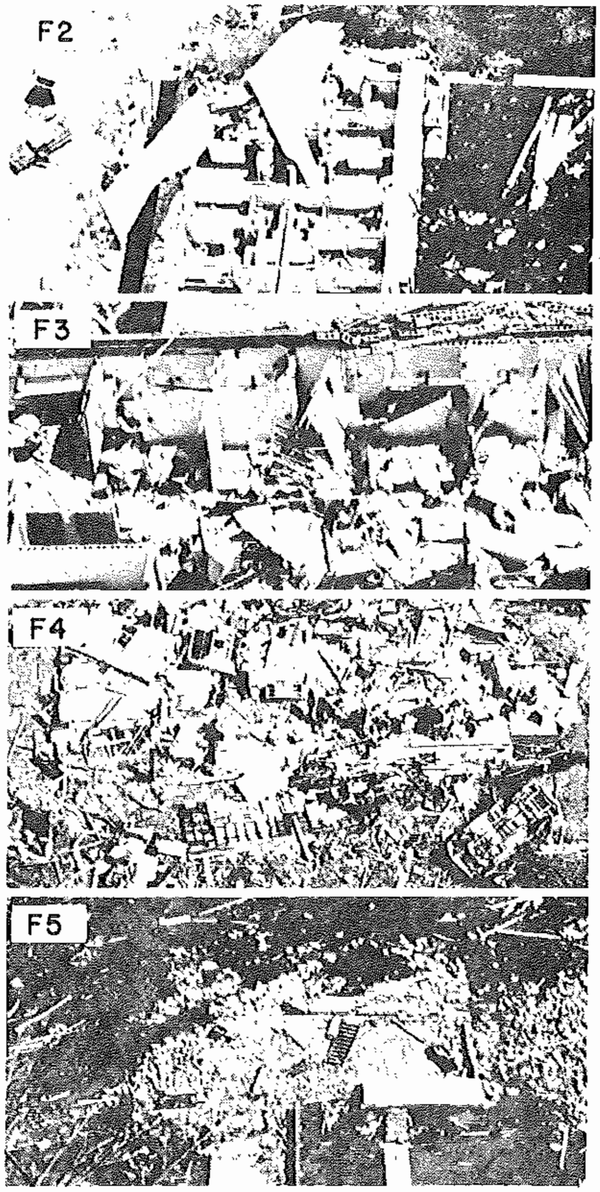


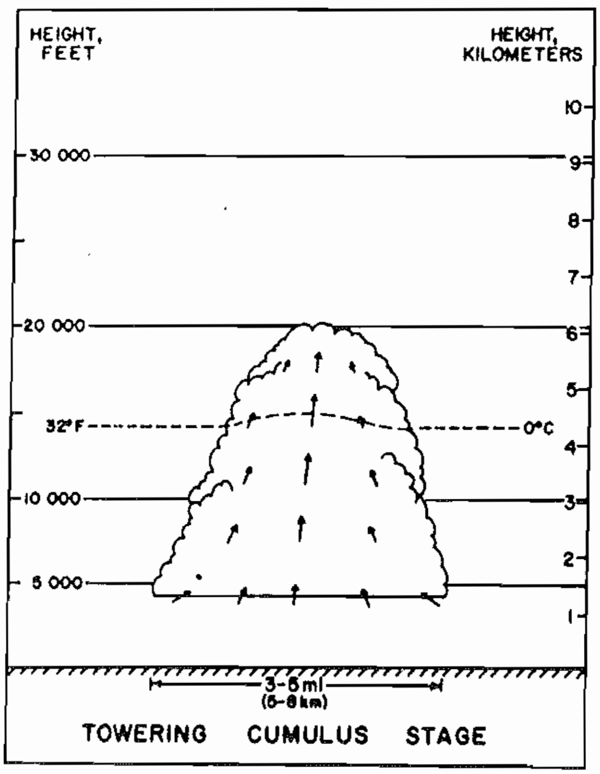
Figure 1-4. Definition and pictorial examples of the Fujita Tornado Scale (F Scale).

Chapter 2
SEVERE THUNDERSTORM MODELS

As mentioned in Chapter 1, severe thunderstorms are the product of an evolutionary process which begins with a single, non-severe cell. From there the structure most often takes the form of the multicell severe and less frequently the potentially deadly supercell. On very rare occasions the storm may remain a single cell and transition into the "pulse-type" storm, discussed at the end of this chapter.

Single Cell, Non-Severe Thunderstorms

The first stage is that of a single cell, non-severe thunderstorm. For a storm to remain in this stage is a rare event. It would be most typical of the short-lived summertime thunderstorm that builds and dissipates generally in the period of 30 to 45 minutes. This storm typically has a three-stage life cycle.



'gure 2-1. In its beginning stage, vertical motion is pre-ominately upward.

During the first (towering cumulus) stage it ingests warm, moist air non-symmetrically from miles around, then lifts and cools it until it condenses (Figure 2-1). A cloud is formed that continues to grow beyond the freezing level. In this stage, vertical motion in the storm is predominantly upward, although small-scale downdrafts can also exist through the process of entrainment (Doswell, 1982). Visually, the upper portion of this storm changes from a fuzzy, soft and feathery appearance to more sharply defined and hard-looking. Near the end of this stage, the first radar echo becomes visible.

When the cloud droplets being borne by the updrafts grow too large and heavy to be sustained, the rain (possibly mixed with small hail) falls back to earth. Momentum from the falling rain is imparted to the neighboring air, forming the downdraft. The storm has now entered the second, or mature stage. It is at this time that the ingredients are created which constitute a "thunderstorm." They are wind, rain, hail, lightning and thunder.

The storm in its mature stage (Figure 2-2) has very sharply-defined upper portions that resemble the look of cauliflower. Both upward and downward vertical motion are taking place within the storm; turbulence can easily reach the severe level. The storm remains a machine that uses warm, moist air for its fuel, and exhausts the cool outflow of wind, rain and hail.

Eventually, in the case of the non-severe storm, the cool outflow spreads out in advance of the storm and undercuts the warm inflow. This immediately disrupts and eventually shuts off the fuel supply to the storm and results in rapid dissipation. As the warm inflow into the storm ends, the third, or dissipating stage begins (Figure 2-3). During this time, the edges and tops of the storm again take on a soft, wispy, feathery appearance as all upward vertical motion has ended and only downward motion is present. The storm becomes more transparent

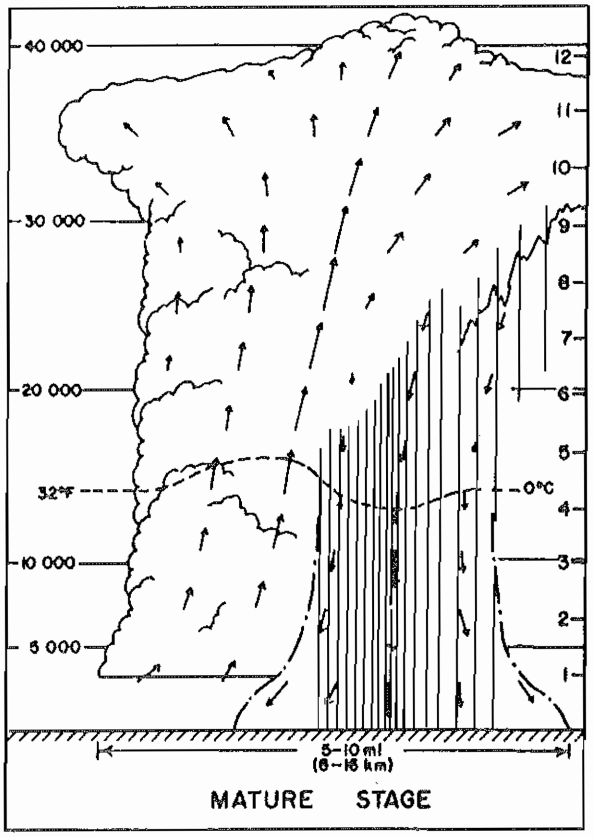


Figure 2-2. It is in the second, or mature, stage that rain, wind and hail may be produced.

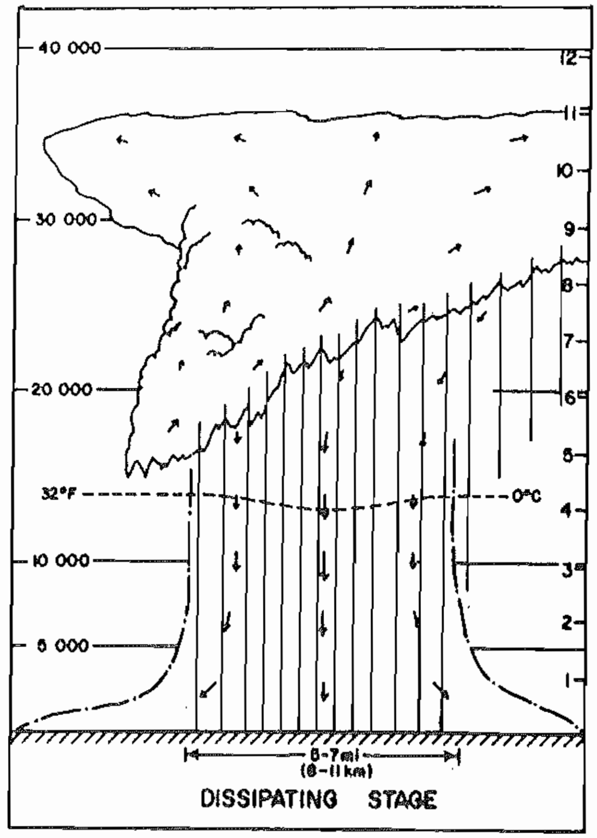


Figure 2-3. The third, or dissipating, stage begins when warm inflow into the storm ends.

with time, eventually disappearing as the last of the rain and/or hail falls to the ground.

The severity of a building storm is dependent on both the vertical shear and stability of the air mass in which it develops. The single cell, non-severe storm can occur in any type of vertical (both speed and directional) shear from strong to weak, since the airmass is usually only weakly unstable (Lifted Indices vary from 0 to -3). This thunderstorm would be poorly organized, with a weak and forward-leaning updraft, and only a weak downdraft.

Multicell, Non-Severe Thunderstorms

Most thunderstorms will evolve into this phase. In the most common situation, the newer cells would build along the southwest flank of the complex while the older cells were dying on the northeast side, as shown in Figure 4. The storm could develop in any kind of vertical shear, but in a weakly unstable air mass with Lifted Indices of 0 to -3. The storm would consist of several updrafts with the individual cells progressing simultaneously through any of the three various stages (towering cumulus, mature or dissipating). Thus, the storm would be only slightly better organized, with moderate and forward-leaning updrafts and moderate downdrafts.

THE SQUALL LINE

The transition in this discussion between non-severe and severe storms is probably best made using the squall line. The squall line begins as some form of convergence develops along an elongated line (Hare, 1986). There are several possible sources of this type of convergence, some of which include: the dry line, orographically-induced lifting, simultaneous lifting associated with an upper level short wave that is only weakly reflected at the surface, or forced lift by a surface front, or mesoscale thunderstorm outflow boundary. The role of lifting, as will be discussed more fully in Chapter 3, is generally to release instability that has built up in the atmosphere, thus triggering the development of the system. With sufficient lifting taking place along this line, convection develops just to the rear of the lift (Figure 2-5). A cool outflow of rain and wind in the exhaust from these storms will fall behind the area of convergence and lifting. In this manner, the warm inflow can continue to lift up over the cool outflow in an undisturbed manner. Once generated, the line can sustain itself through continued low-level convergence. It will continue to exist until it has moved into an area where warm, moist convergence into the system no longer is available.

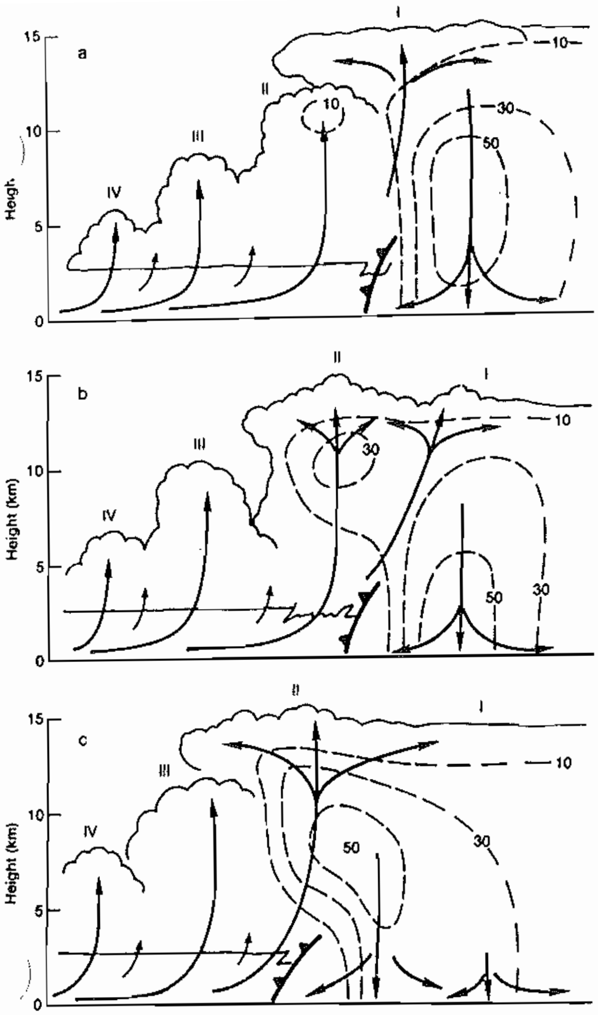


Figure 2-4. Multicell storms are composed of cells in various stages of development. Cells, as depicted above, take turns in being dominant.

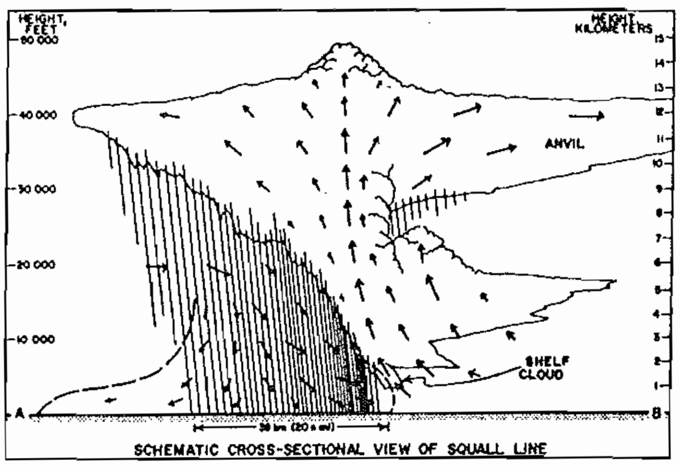


Figure 2-5. Warm air is lifted along the leading edge of the cool squall line outflow, producing rain along and behind the outflow.

Squall lines are of two types. The first consists of closely spaced cells with nearly continuous precipitation (Figure 2-6). This type rarely produces severe weather. The second type is composed of more isolated cells within the line. Severe weather is more likely to occur with this second type, since individual cells do not have to compete for moisture and upward motion (Figure 2-7). The distinction as whether a line will consist of isolated cells or be more continuous appears to be related to vertical wind shear and the strength of the low-level inflow. It is not uncommon for a line of isolated cells to gradually evolve into a more or less solid line by progressive filling of gaps. When tornadic activity is associated with isolated cells, the end of such activity usually is signalled by this so-called "lining out" of the cells (Doswell, 1982).

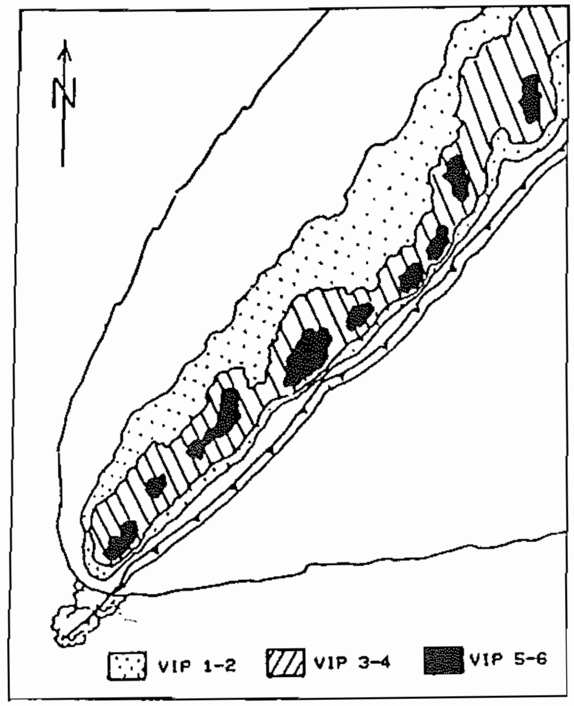


Figure 2-6. The continuous squall line is rarely a producer of severe weather.

The squall line is a well-organized system in which warm, moist inflowing air is lifted up over much cooler and moist outflowing air. This separation of flow regimes prevents outflow from interfering with inflow. As a result, needed fuel for the system will remain available, making an extended lifespan possible.

The atmosphere in which the squall line commonly forms is characterized by strong directional shear in the low levels (from the surface to 700 mb) with little or no shear above 700 mb. Typical Lifted Index values may vary from -2 to -5. The storm system develops moderate updrafts and moderate to strong downdrafts.

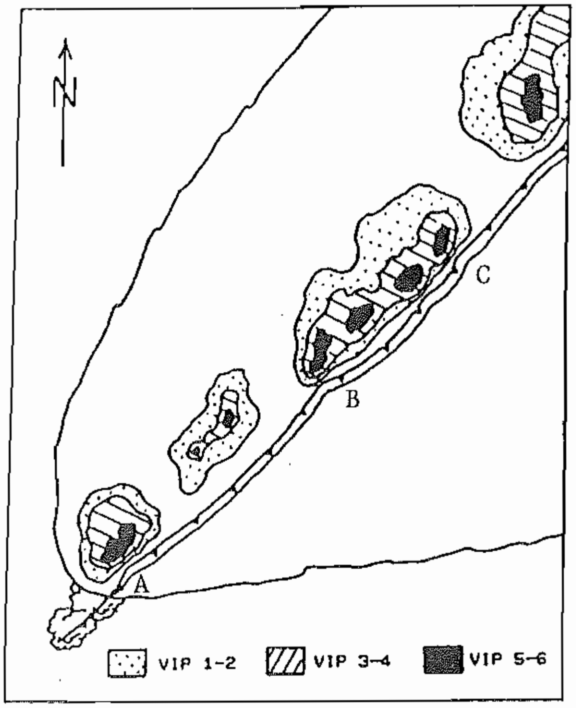


Figure 2-7. Severe weather is more likely to occur in a broken, rather than a continuous, squall line. Preferred areas for severe weather would be at points A, B, and C.

The most frequent form of severe weather associated with a squall line is damaging wind from precipitation-induced downdrafts and the downward transport of horizontal momentum. Severe weather that does occur usually meets only the minimum criteria (wind gusts to 60 mph, hail of 3/4 inch diameter or isolated weak tornadoes.) The preferred area for severe weather would be where low level moist inflow is not obstructed, such as the inflow end of a line, upwind side of a break in the line, or at the crest of an LEWP. Lines are usually oriented north-south, northwest-southeast, or northeast-southwest. In each case, severe weather would be most likely on the south end of the line or north of a break in the line.

Multicell Severe Thunderstorms

Oftentimes multicellular thunderstorms can become severe. Tornadoes are possible, but are usually weak or minimal in strength, with the main severe weather threat being large hail and damaging wind. Extreme low-level shear associated with these damaging winds, often referred to as "downbursts", have been responsible for numerous notable aircraft accidents over the past ten years.

DOWNBURST

The downburst is classified on the basis of spatial scale. A macroburst affects a circular area 2.5 miles in diameter or larger, whereas a microburst affects a circular area less than 2.5 miles in diameter. Most damaging events are microbursts, or combinations of several microbursts.

Microbursts can be further separated into two types: wet or dry. It should be noted from the outset that all reported microbursts have been associated with storms that produced rainfall. However, the distinguishing factor as to whether to classify a particular microburst as dry or wet is based on the amount of precipitation reaching the earth's surface. In general, using suggestions by Wakimoto (1985), the events are defined as follows.

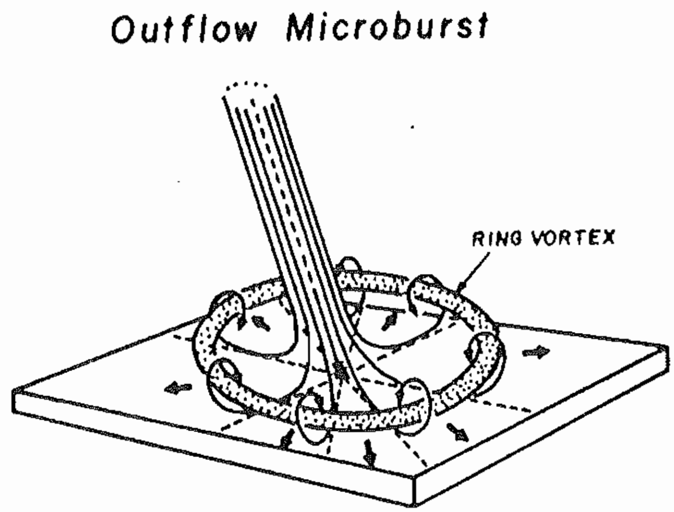


Figure 2-8. Microbursts strike the surface, producing asymmetric flow in all directions (Fujita, 1985).

Dry microbursts are accompanied by little or no rain (<0.01 inches) at the surface between the onset and end of high wind, including intermediate calm periods.

Wet microbursts are accompanied by measurable, oftentimes heavy, rain at the surface between onset and end of high wind, including intermediate calm periods.

Much work has been done on the dry microburst (Figure 2-9) and consequently much is known regarding the environmental conditions under which it occurs and the mechanisms which force it. Apparently, the major driving force behind the dry microburst is negative buoyancy generated by extensive evaporative cooling within a deep dry adiabatic subcloud layer. Such conditions are favored by altocumulus or shallow, high-based cumulonimbus with virga emanating from their bases. Dry microbursts occur with great regularity during the warm months over the Western Plateau of the United States.

The wet microburst (Figure 2-10), on the other hand, has been receiving increasing attention, especially in the aftermath of the Pan Am Flight 759 (New Orleans) crash in 1982 (Fujita, 1983) and the Delta Flight 191 (Dallas-Ft

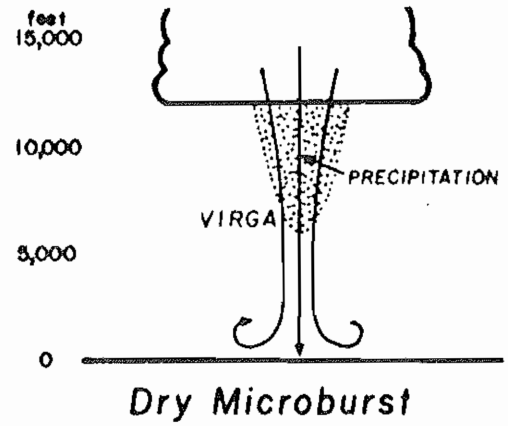


Figure 2-9. Dry microbursts are accompanied by little or no rain (Fujita, 1985). Outflow is generally produced by evaporative cooling.

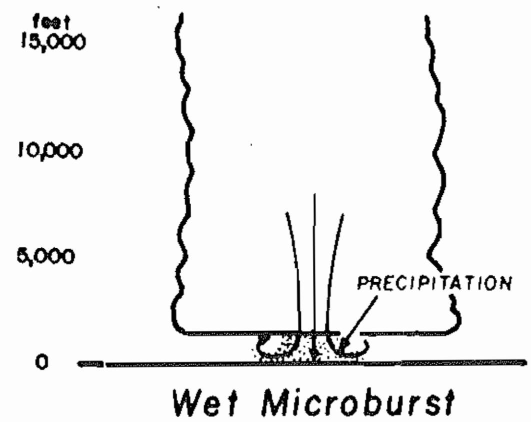


Figure 2-10. Wet microbursts are accompainied by measureable rain at the surface (Fujita, 1985). Outflow is produced by a combination of evaporative cooling and precipitation loading.

Worth) disaster in 1985 (Caracena et al, 1986). These events led to a series of comprehensive field experiments in the Southeast region of the U.S. (Dodge, 1986). Based on these experiments and disaster analyses, the wet microburst is thought to develop within an environment characterized by a near moist-adiabatic subcloud lapse rate and deep low-level moisture. The forcing mechanism appears to be a combination of subcloud evaporative cooling and precipitation loading from the abundant liquid water content of the clouds.

Microbursts are about equally likely to produce a warming effect, due to adiabatic compression, as a cooling effect. Bedard (1987), has demonstrated analyses of microbursts in which sufficient warming occurred to produce enough positive buoyancy within the microburst to prevent impact with the surface. Furthermore, because of the small spatial scale, microbursts can strike the ground between wind sensors and go undetected. This means that microbursts which can severely affect aircraft take-offs and landings may not be detectable from surface data. Bedard suggests, however, that a characteristic pressure nose may be associated with each microburst that, if noted, could allow warning times from a few seconds to as much as one minute. Additionally, doppler radar, to be installed at major airports in the 1990s, should greatly enhance the detection of microbursts.

In a study of downbursts that occurred in north Texas through the year 1985, Read et al. noted that the events showed several similarities. In each case, the parent thunderstorm formed on the warm side of a pre-existing thermal boundary. In most cases, the boundaries had been produced by earlier thunderstorms. Moisture convergence increased in the region of formation, resulting in a pooling of higher dew points near the boundary. For an explanation of moisture convergence, see the Analysis Information section in Chapter 4. Unexpectedly, the storms developed under conditions in which temperatures were not unusually high. In fact, most of the trigger temperatures were at or below the normal climatological maximum for the date of occurrence. This was possible because little or no capping inversion was present. An in-depth discussion of the importance of this capping inversion (also known as the "lid") will follow in Chapter 3.

Thunderstorm bases on these days were near the Convective Condensation Level (CCL), which was generally at or above 5000 ft. The atmosphere approached dry adiabatic conditions from the surface to the CCL. A moist mid-level layer existed from 5000 to 15,000 ft., with another dry layer above 15,000 ft. It is believed the thunderstorm downburst was accelerated by three factors. First, water droplets within the thunderstorm downdrafts could have been chilled as they encountered dry air in the subcloud layer. This effect, called evaporative cooling, would have increased both the weight of the air in the outflow and its speed of descent (increased negative buoyancy). Secondly, high moisture content was indicated on Skew-T plots adapted to the thunderstorm generation region. Thus, great amounts of water could have been lifted upward by the thunderstorm updrafts. As the weight of moisture becomes too much to be sustained by the updraft, the rain descends to the ground with the wind, increasing its speed. This effect, in which raindrops impart momentum through friction to the surrounding air, is called "precipitation loading". Finally, momentum from strong winds aloft could have been transported downward to the surface in the dry air of the subcloud layer. This vertical transport of horizontal momentum, in which air with higher momentum aloft is mixed to progressively lower levels, raises the momentum of the air in the lower levels.

Microburst-producing thunderstorms generally formed in an atmosphere that shows weak vertical shear, but very strong instability (including large positive lifting areas on Skew-T charts). Lifted Indices are between -5 and -10. These storms are well-organized, with moderate updrafts and very strong downdrafts. Several models of multicellular severe thunderstorms have been developed. The most prominent of those within the literature are listed below.

LINE ECHO WAVE PATTERN

The Line Echo Wave Pattern (LEWP) is a feature associated with the more severe squall lines. It develops as one segment of a squall line begins to accelerate forward and forms a bulge in the line. This is a direct result of an increasing pressure gradient brought about by a concentrated burst of cold downdraft air impacting the surface and forming a mesohigh immediately to the rear of the bulge. As the forward acceleration continues, pressures lower in the trailing segment of the line, often forming a mesolow. The resulting configuration of a forward bulge and a trailing crest makes up the characteristic signature of the LEWP. This bulge in the LEWP can result in the formation of a "bow echo", a feature commonly associated with downbursts (Fujita, 1978). This will be discussed in greater detail later in the chapter. In addition to the potential for downbursts, tornadoes occasionally develop on the north side of the bulge where cyclonic shear predominates. With the first evidence of LEWP formation, the forecaster should realize that the potential for severe weather has greatly increased.

BOW ECHO

An occasional form of the multicellular severe storm is the bow echo (Fujita, 1978). This phenomenon has a three-stage life, as shown in Figure 2-11. In its initial stage it develops as a short, narrow line of thunderstorms, perhaps evolving from a Line Echo Wave Pattern (LEWP). As it begins its second stage, the downburst phase, one of the cells in the line surges forward, creating the "bow" in the line. Near this time, the cell's top collapses and a strong downburst is frequently produced. Often a low-level echo weakness forms on each side of the "bow" (due to subsidence associated with the downburst.) In its final (or dissipation) stage, the echo takes on the shape of a comma. Usually, cyclonic shear near the head of the comma will produce rotation. If this rotation occurs in a favorable environment, a tornado may occur.

The bow echo develops in an environment of high moisture content and strong instability, with Lifted Indices generally between -5 and -9. There is usually a layer of dry air present at the mid levels. It produces its damaging winds by a combination of evaporative cooling and water loading. Momentum transport may also play a part in the production of damaging winds, since both a low- and upper-level jet are usually present. It is thought that the coupling of the two may assist in the momentum transfer (Ucellini, 1979).

DERECHO

An even more rare form of the multicellular severe storm is the "derecho", which is a family of downburst clusters. This term was first introduced by Hinrichs in 1888. The first type of derecho is usually a winter or spring event, associated with a strong extra-tropical low pressure center. It usually develops as an accelerating portion of an extensive squall line. The second, and more frequent, form of derecho occurs in the Midwest in the late spring or summer, generally associated with weak synoptic scale systems.

The derecho can encompass a span of several hundred miles (Johns, 1983, 1986, 1987). The progressive version of the derecho, as shown in Figure 2-12, accounts for about three-quarters of the events. It appears like a short,

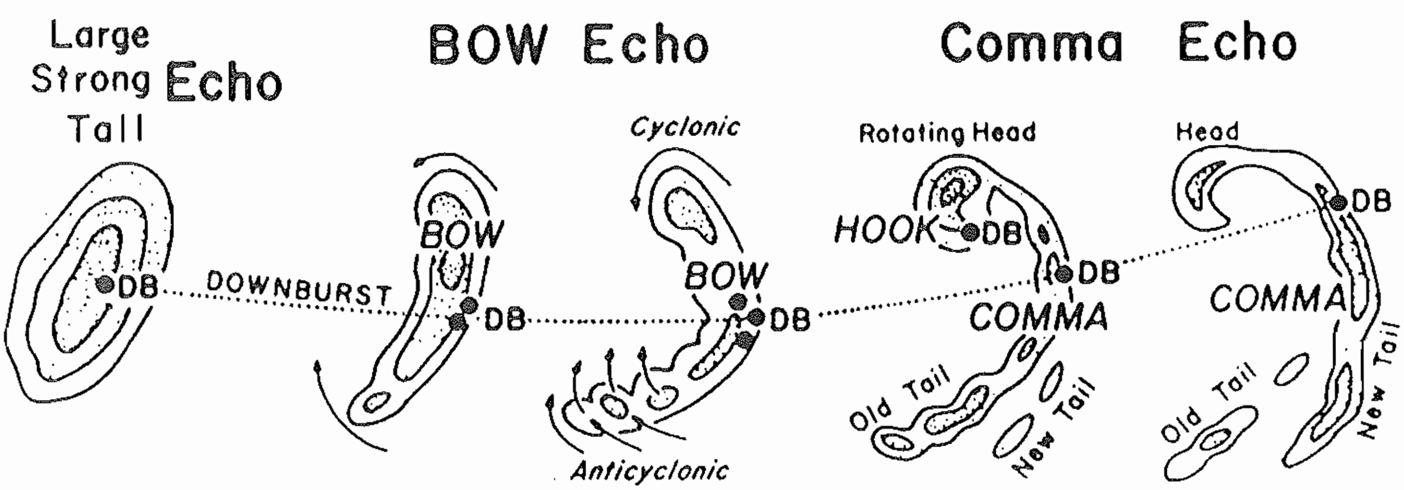


Figure 2-11. The bow echo develops as a short, narrow line of thunderstorms. One cell in the line produces a downburst (point DB) and the system takes on the bow shape. In its dissipating stage, the echo takes on a comma shape. (After Fujita, 1978).

curved squall line on radar, oriented nearly perpendicular to the mean windflow. Downburst activity generally occurs near a bulge that forms in the line. This configuration may resemble a bow echo at times. At other times, an eastward extension may form downwind from the north end of the line, giving the appearance of an LEWP. The configuration may, in fact, shift its appearance between the two. Severe weather in the form of damaging wind usually occurs near a bulge in the line; however, this bulge can be a zone from 50 to 180 nm in length. Isolated tornadoes occasionally accompany the system.

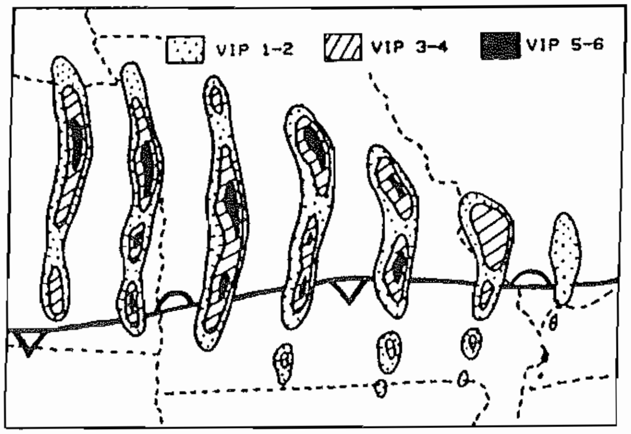


Figure 2-12. The derecho is a long-lived line of thunderstorms that consists of evolving line echo wave patterns and bow echoes.

The serial pattern of the derecho occurs nearly one-quarter of the time. It appears like a squall line oriented in nearly the same direction as the mean wind flow. Damaging wind occurs with a series of LEWPs and bow echoes that move along the line.

The derecho-producing system generally begins just to the north of a quasi-stationary east-to-west frontal boundary, in a zone of very strong low-level warm advection and deep moisture. Surface dew points are generally high, suggesting that pooling of moisture through convergence helps to set off the activity by increasing the instability and lowering the Level of Free Convection (LFC). Weak warm advection at 700 mb assists in producing upward vertical motion. The atmosphere is very unstable, with Lifted Indices between -6 and -10. Strong vertical speed shear is usually present, with a weak short wave moving over the area and a primary jet stream well to the north.

Once initiated, the system tends to propagate more rapidly than other activity in the area, usually at greater than 5 knots. It continues to produce downburst activity for many hours and travels for several hundred miles. It generally moves to the right of the mean wind vector and at a faster speed than the mean wind. This would suggest a movement to the southeast, and eventually into the warmer air mass.

Supercell Thunderstorms

The supercell (Figure 2-14) is the last transitional phase in the evolution of a severe storm. It is potentially the most devastating phase, capable of producing simultaneously the entire array of severe weather: large hail, extreme winds and violent tornadoes. While still in the multicellular phase, one updraft becomes dominant. This updraft, which had been forward-tilted, intensifies. Consequently, it is affected less by environmental winds and becomes more vertical. An organized downdraft area develops immediately downstream from the intensifying updraft, thus separating the rear intake from the forward exhaust and eliminating any interference between the two (Figure 2-13). The rear portion of the thunderstorm becomes rain-free as inflow and upward motion increase.

Continued development is marked by the eventual formation of a Weak Echo Region (WER). At this time, the updraft continues to intensify along the low-level inflow side of the storm (usually the southeast). As the intensity of the updraft increases, it begins to rotate. This rotation may be a result of adjustments of the updraft to satisfy an extreme vertical pressure gradient. It appears that only a rotating updraft is efficient enough to reach the speed required. Eventually the updraft becomes so intense that air parcels are lifted upward so quickly they do not have time to condense until they reach the mid levels of the storm. This produces a WER beneath a mid-level overhang. Further increase in updraft speed can result in an

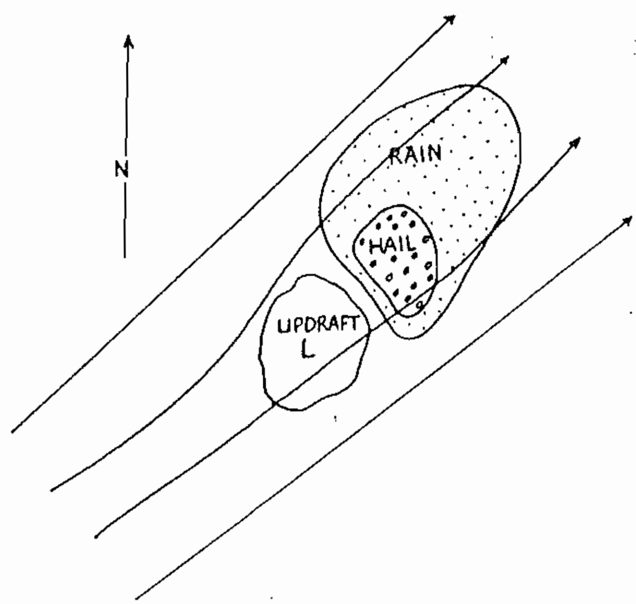


Figure 2-13. As a thunderstorm becomes better organized, it develops a separate rear inflow into its updraft and forward exhaust of wind, rain and hail.

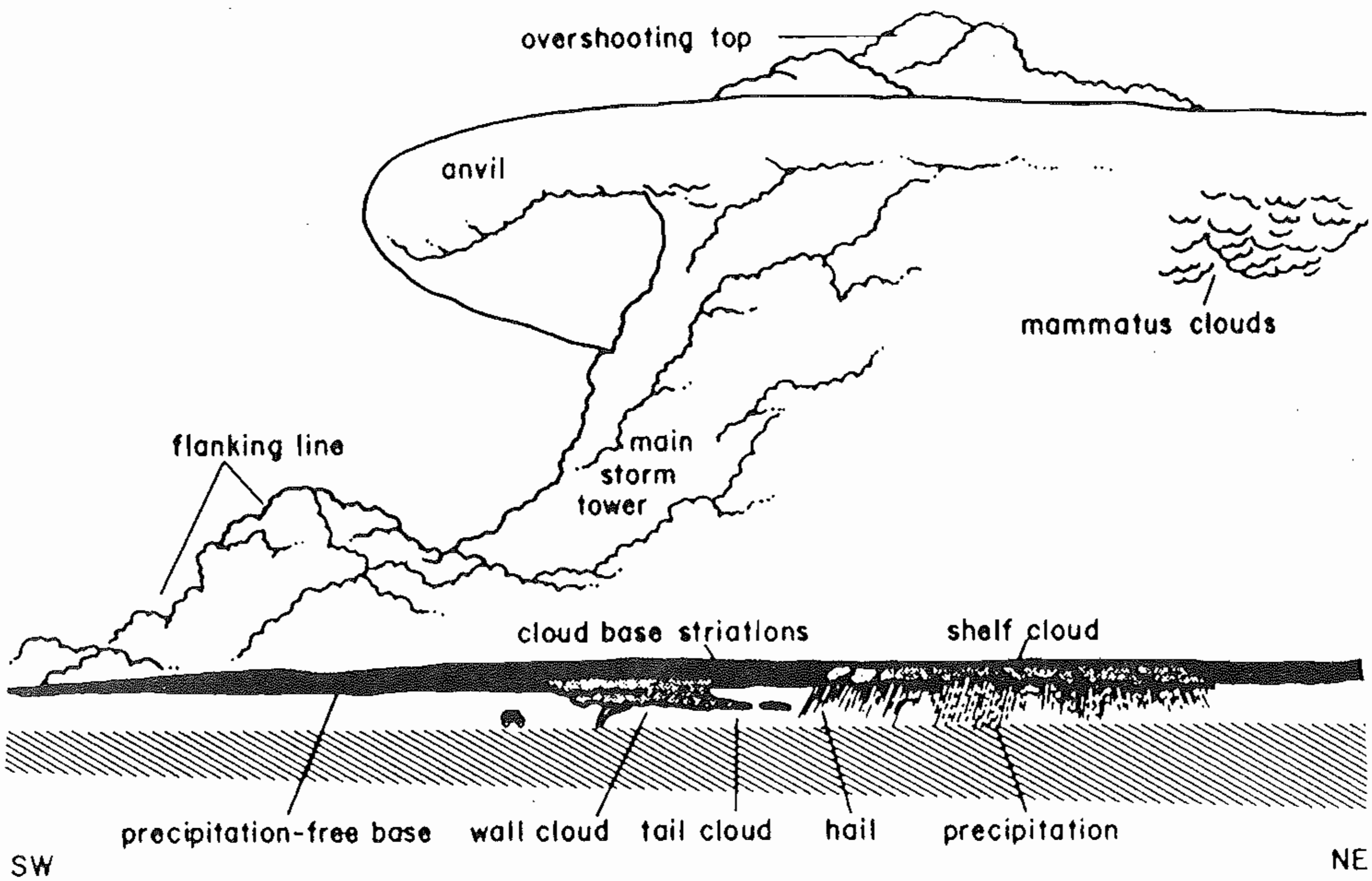


Figure 2-14. Side view of the Supercell thunderstorm. The storm is depicted as moving from left to right. Outflow (in the form of wind, rain and/or hail) occurs in the right, or front portion. The inflow to the storm is at the left, as indicated by the smooth, rain-free base. A tornado would be most likely to occur under the rain-free base, in the area of the wall cloud. (Diagram by C. Doswell and B. Dirham).

echo-free updraft in both the lower and mid levels. This area, surrounded by radar echoes, is called a Bounded Weak Echo Region (BWER). After the BWER forms, a mesocyclone develops in the mid levels in the area of the BWER. The lowering of the rain-free base called the wall cloud usually takes place at this time.

As described by Lemon and Doswell (1979), the rotation of the storm's updraft eventually becomes so intense that the environmental winds can no longer move through the updraft. It is now acting like a solid cylinder in a stream. The mid-level winds must flow around it on either side, and the winds striking the updraft on the back side have no choice but to descend to the ground (Figures 2-15 and 2-16). This is referred to by many researchers (Barnes, 1976; Nelson, 1977; Lemon and Doswell, 1979) as the "Rear Flank Downdraft" (RFD). It is often observed at this time that the storm is no longer being driven by the mid- and upper-level winds, but slows and begins to move to the right (on rare occasions, to the left) of its previous track.
There are several theories as to why this occurs. The first is that the rotation of the updraft and mid-level winds are in the same direction and add together on the right side, but are in opposite directions and subtract from each other on the left side. In accordance with the Bernoulli principle, a faster net wind and lower pressure develop on the right side. This, in turn, sets up a pressure gradient that moves the storm to the right. Another widely proposed theory is that the storm simply moves in such a way as to maximize the low-level feed of warm moisture, i.e., generally to the right or southeast. This is the reason for the warning that a storm that suddenly changes its speed or direction may be developing into a severe storm.

The RFD is accelerated downward (Figure 2-17) by the evaporative cooling of precipitation in the mid levels of the storm. As it descends, a pendant echo often is noted on radar. Continuing downward, the mesocyclone becomes displaced from the BWER or updraft to a location upwind, between the BWER and the RFD. This mesocyclone becomes a second rotating updraft Figure 2-18). It is located in a zone of very strong vertical shear, between the original warm updraft (or BWER) to the east (in the

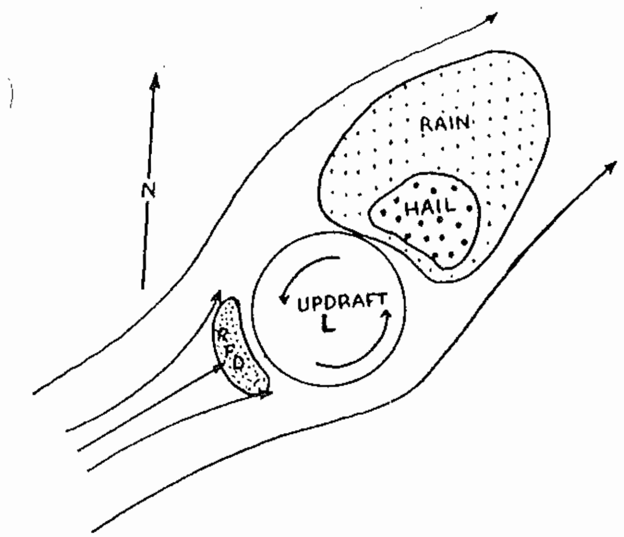


Figure 2-15. Environmental winds striking the strengthened updraft are forced earthward, producing the rear flank downdraft (RFD).

warm inflow sector) and the cool Rear Flank Downdraft to the west. At this time, if a wall cloud is present, it will often begin to rotate.

The final stage of the supercell life is the BWER collapse stage. The separation of the mesocyclone from the BWER and the formation of a second updraft in the storm causes the main updraft, or BWER to weaken. As the RFD finally strikes the surface, it creates a second outflow boundary in the rear of the storm. This creates a wave feature in the mesoscale pressure field of the storm. The forward projection of this rear outflow causes the wave to occlude, decreasing the warm inflow into the main updraft and weakening the updraft further. At this time, the storm top lowers as both the BWER and the mesocyclone descend to the surface. A tornado may then descend toward the surface in the area of the mesocyclone (Figure 2-19). The tornado generally exhibits a two-stage life cycle.

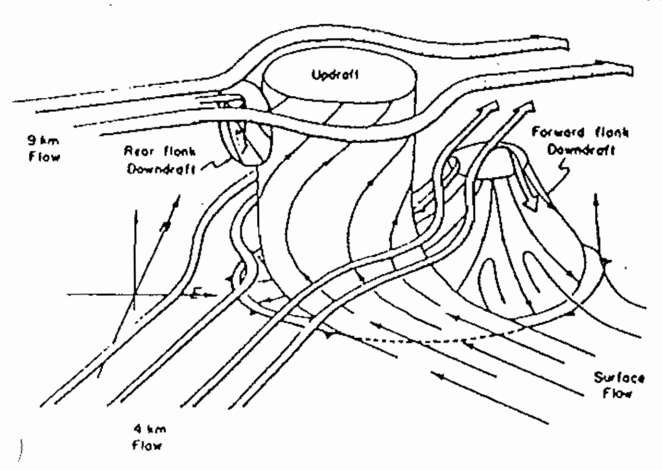


Figure 2-16. Side view of the developing Rear Flank Downdraft (Lemon and Doswell, 1981).

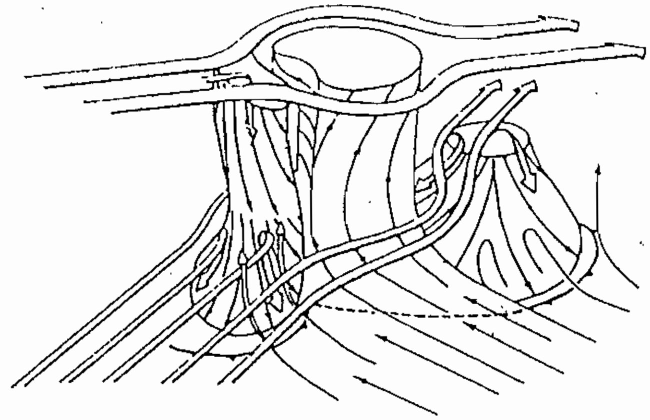


Figure 2-17. As the RFD strikes the surface, it produces an outflow that begins to wrap around the updraft. (Lemon and Doswell, 1981).

First, a funnel forms under the rain-free base of the storm (in the area of the wall cloud if one is present.) The funnel then lowers toward the surface. Often, damage is seen rising from the ground even before it is obvious that the funnel has reached the surface. In such a case, a tornado is occurring, and the connection will become visible as soon as sufficient condensation occurs or enough damage debris has been lifted into the tornado. As the tornado progresses through the first stage, it becomes wider and eventually reaches its maximum size. As the surface wave fully occludes, the RFD wraps around the tornado and cuts off the warm inflow into the parent mesocyclone. The tornado now starts its dissipating phase. It shrinks back into a thin rope and becomes tilted. Finally, it lifts back into a funnel and disappears.

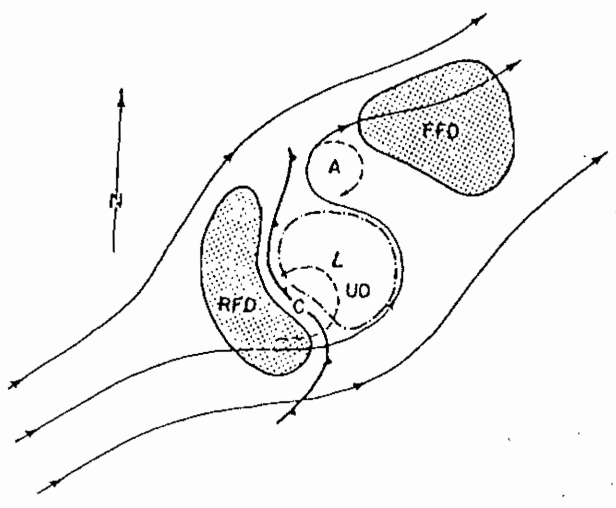


Figure 2-18. Storm-relative flow at about 20,000 feet. The original updraft is labeled L while the new mesocyclone is at C. (Lemon and Doswell, 1981).

For the moment, the period of severe weather is over. If warm inflow redevelops into the occluded low at the surface, a main updraft can regenerate over the inflow and the entire cycle of development and severe weather production can occur again. This is depicted in Figure 2-20.

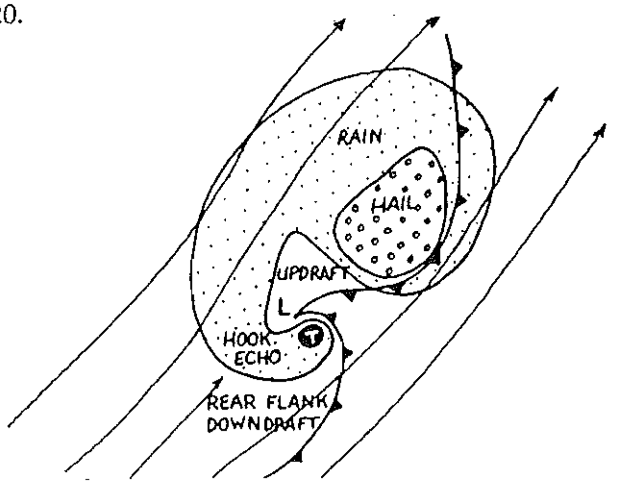


Figure 2-19. Approximate location of tornado as RFD is wrapping around the storm main updraft.

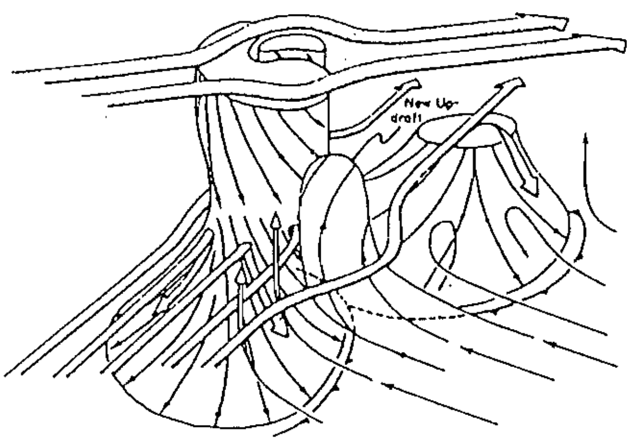


Figure 2-20. As the RFD occludes, the original updraft weakens. A new updraft can form southeast of the original updraft (Lemon and Doswell, 1981), reintroducing the possibility of severe weather.

Other Forms of Severe Thunderstorms

The following subclasses of severe storms occur much less frequently than do the bow echo, derecho or supercell. For this reason, they are sometimes referred to as "secondary classes."

PULSE THUNDERSTORM

The pulse thunderstorm was first described by Chisolm and Renick in 1972 as a "single cell hailstorm." The pulse storm (Figure 2-21) is very well-organized, with moderate vertical or forward-tilted updrafts and intense but very short-lived downdrafts (Doswell, 1985). It exhibits elevated bases, with the first radar images often above 20,000 ft (Wilk et al., 1979), and usually builds quickly to VIP 5. Should the VIP 5 portion of the storm reach as high as 30,000 ft, the probability of large hail is greater than 50%. The intense, elevated core of the storm then descends rapidly to the surface, producing severe weather at the same time. Again, it is most likely to produce large hail or damaging winds, but rarely produces tornadoes.

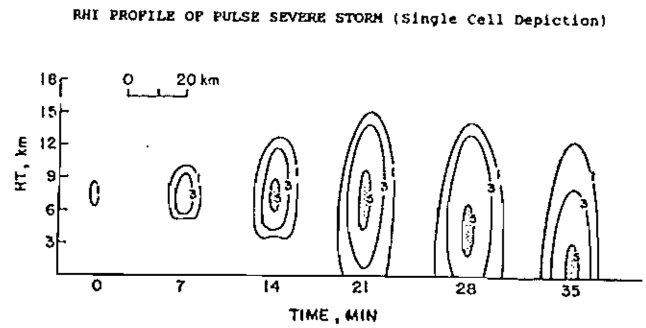


Figure 2-21. Pulse storm (after Wilk, 1979).

DRYLINE THUNDERSTORM

This storm (so named because it forms along the dryline) was only recently recognized as a separate class of severe storm. It exhibits a strong, rotating updraft, much like the supercell storm, but can have a deceptively weak appearance on radar (Doswell, 1982). This is because it is a very inefficient rain producer, forming only a weak precipitation core. However, the precipitation it does produce often includes large hail. Mid-level flow (relative to the storm) is also weak and it takes in very little dry mid-level air. With neither water loading nor evaporative cooling to assist, it does not form strong downdrafts. It can produce damaging tornadoes.

As the storm matures, as depicted in Figure 2-22, it increases the rotation of the updraft while the size of the updraft in the mid levels begins to decrease. This greatly limits the amount of warm inflow available to the storm. Eventually the mid-level portion of the storm collapses and the storm dies.

On radar, this storm is again similar to the supercell, with strong reflectivity gradients on its inflow flank. The most striking characteristic, however, is its visual appearance. Perhaps because of the weak moisture amounts in the low levels, it often has no accessory clouds, such as roll and shelf clouds. So, without dense precipitation or accessory clouds in the way, the rotating updraft becomes visible. Often striations are present that show the spiraling action of the updraft even more markedly.

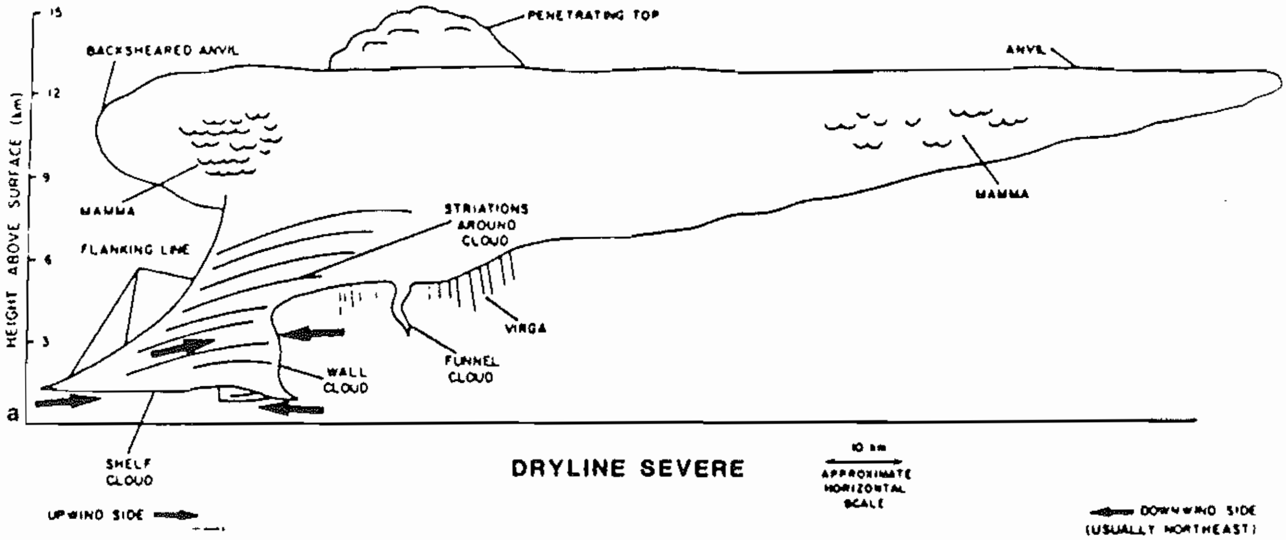


Figure 2-22. With limited moisture available for development, the dryline severe thunderstorm produces little rainfall and may present a deceptively weak radar appearance.

SUMMARY

A spectrum of thunderstorm types exists through which storms evolve. They first develop as single (non-severe) cells, with many later building into multiple (non-severe) cell clusters. From that point the storms that will become severe make the transition during the multicell portion of their life. They do this by calling upon a combination of instability and/or vertical shear to intensify their updraft strength. Finally, if both great instability and great vertical shear are available, a very few storms will intensify to the supercell stage. Figure 2-23 below summarizes the relationship between vertical shear and updraft strength. Figure 2-24 indicates generalized vertical winds profiles associated with different thunderstorm classes.

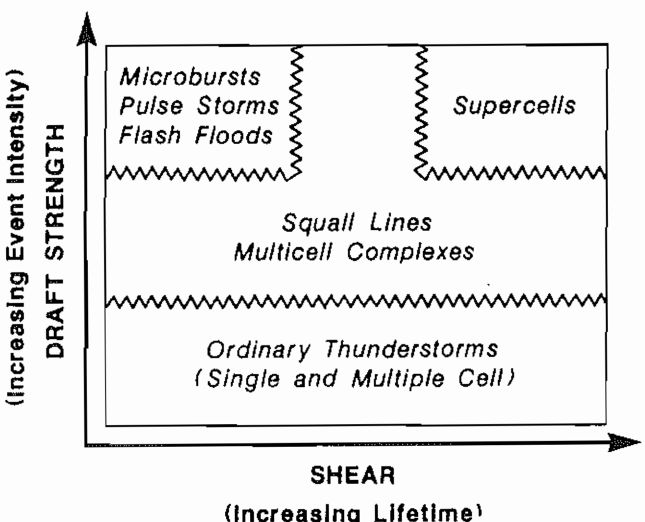


Figure 2-23. Relationship between updraft strength and shear in the development of severe thunderstorms. (Moller and Doswell, 1988).

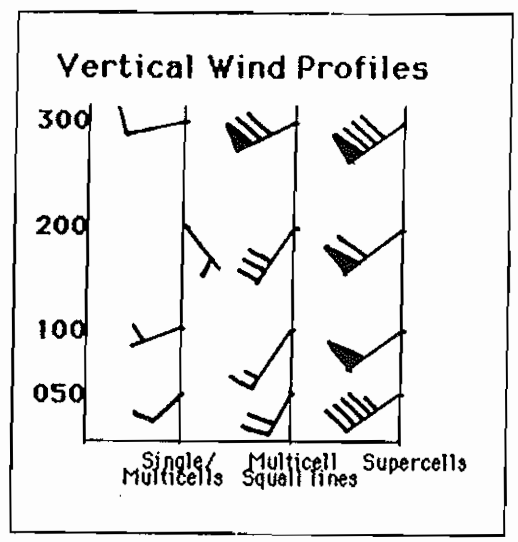


Figure 2-24. Typical vertical shear regimes associated with thunderstorms.

Chapter 3

PRE-EVENT OPERATIONS

This chapter is concerned with the knowledge that the forecaster can use to gauge the potential for severe weather and "get ready" for the event. Specifically, it discusses how various atmospheric parameters come together to differentiate between the non-severe and the severe weather day; and, to some extent, between the day with extreme severe weather and the day with minimal severe weather.

Convection, whether severe or non-severe, requires the presence of three ingredients. They are instability, sufficient moisture and a trigger mechanism. Generally, the degree to which each is present will determine the potential for severe weather.

Instability

Instability can be defined as the tendency of a parcel of air to move vertically through the atmosphere. If a parcel that is initially forced away from its original level moves farther away from that level, the atmosphere is said to be unstable. If, on the other hand, an initial force results in the parcel moving back toward its original level, the atmosphere is stable. Neutral stability can exist in which a force must be applied continually to the parcel to move it. If the force is removed, there is no tendency for the parcel to move back to or farther away from its original level.

In assessing the instability of the atmosphere for a given layer, it is necessary to evaluate how the temperature of the parcel changes with respect to that of the surrounding environment. In other words, we must consider the environmental lapse rate versus the parcel's adiabatic lapse rate.

The temperature of a lifted air parcel will decrease at the dry adiabatic rate as long as it remains unsaturated. Prolonged lifting in an environment that is somewhat moist eventually will produce saturation of the parcel, in which case movement is along a moist adiabat. Regardless of its moisture content, a parcel will be positively buoyant as long as its temperature remains warmer than the environmental temperature.

The path etched out by a positively buoyant parcel on a thermodynamic diagram defines the positive area, and this area is proportional to the amount of kinetic energy the parcel gains from the environment. Conversely, the path taken by a negatively buoyant parcel defines the negative area, which is proportional to the amount of kinetic energy the parcel loses to buoyancy. In other words, energy must be supplied to the parcel (usually through mechanical means) in order to overcome the negative buoyancy.

It is possible, and not infrequent, for a number of these positive and negative areas to exist on a given sounding below 100 mb. How many and to what extent they exist depends on the parcel chosen and whether its movement results from surface heating, release of latent heat of condensation, or forced lifting. The important consideration appears to be the size of the positive areas versus the negative areas on the sounding, i.e. the surplus of parcel energy available.

If we consider the case of forced lifting as shown in Figure 3-1, we see that a surface-based unsaturated parcel is lifted dry adiabatically until it crosses the mixing ratio line corresponding to the surface dew point temperature. This point of intersection is the Lifted Condensation Level (LCL) and is where saturation of the parcel first occurs. From this point, further lift will be along a moist adiabat. However, notice that the parcel is still negatively buoyant and requires additional input of energy through forced lifting to rise further. If this energy is sufficient, the parcel may reach the Level of Free Convection (LFC), at which time it will become positively buoyant. The parcel's speed of ascent increases until it reaches a point where it again becomes cooler than its environment. This point is called the Equilibrium Level (EL).

The EL is rather significant when one considers that the tops of thunderstorm anvils are most frequently

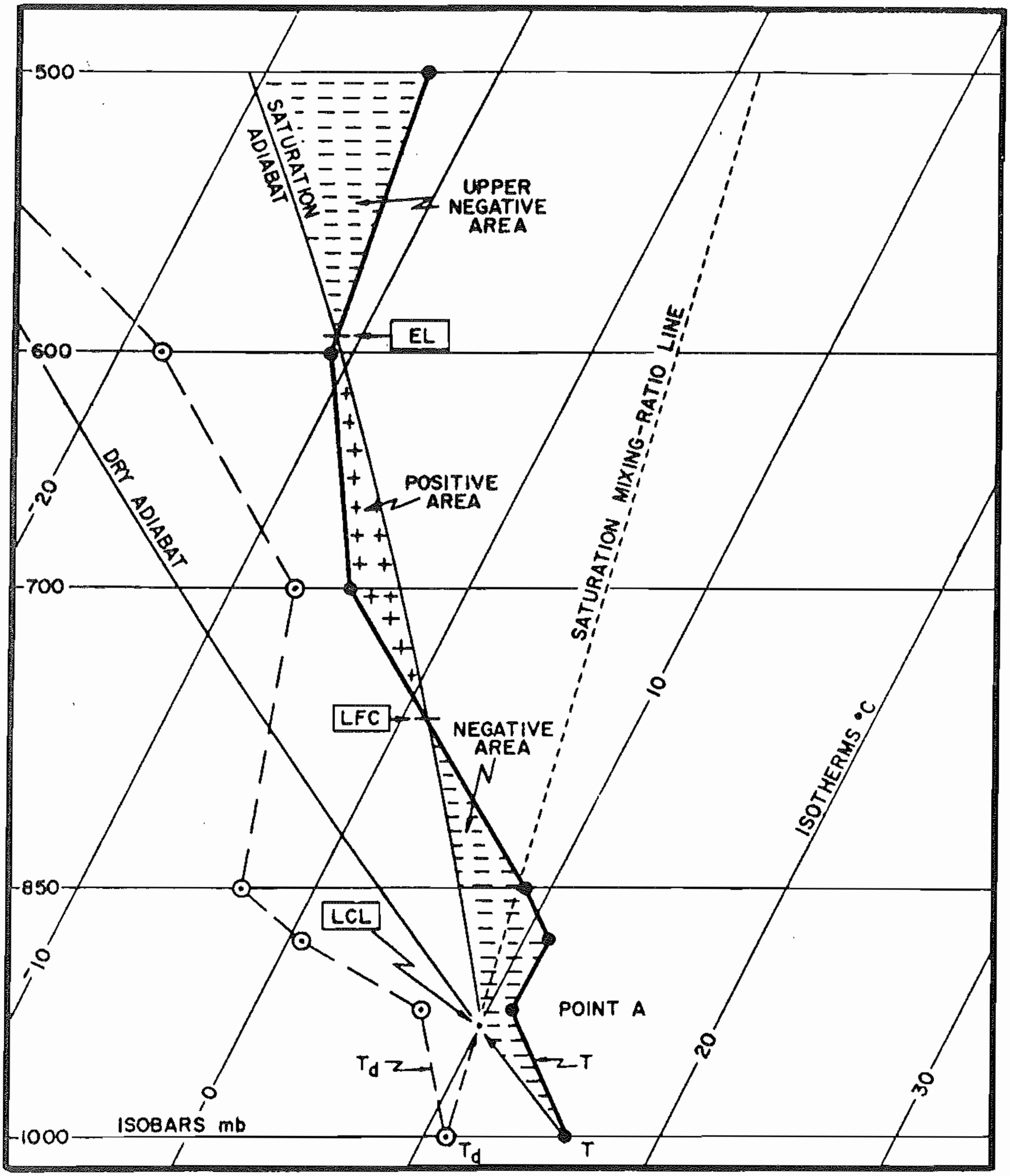


Figure 3-1. Positive and negative areas on a sounding (after AWSM 105-124).

found here and not at the tropopause. A rising parcel has its maximum momentum upon reaching the EL. Above this level, the momentum is damped out rapidly as the parcel enters a region of strong negative buoyancy. However, as long as some momentum is available, "overshooting" of cloud tops will occur. The height of the overshooting top is proportional to the amount of momentum which must be exhausted and is indicative of strong convection. Eventually, a level is reached where the momentum falls to zero. This is referred to as the Maximum Parcel Level (MPL). Graphically, the MPL is placed where the size of the negative area above the EL equals the size of the positive area between the LFC and the EL.

Two special cases of instability often exist in the atmosphere which are of critical importance to the operational meteorologist. Conditional instability exists whenever the environmental lapse rate lies between the dry adiabatic and moist adiabatic lapse rates. Under this condition, the atmosphere is stable for unsaturated parcels and unstable for saturated parcels. Conditional instability is common year-round in the tropics and during the summer months in the middle latitudes.

Convective or potential instability, on the other hand, is a layer concept that is highly dependent on the vertical moisture profile in the atmosphere. Layers in which the wet-bulb temperature lapse rate is greater than the saturation adiabatic rate are said to be convectively unstable. Such conditions result from warm moist air underlying cool dry air. When a layer of air with this structure is lifted, lower level parcels will reach saturation earlier than parcels at the upper portion of the layer. As such, these parcels that follow dry adiabats will cool faster than lower level parcels that are cooling moist adiabatically. The net result is a decrease in the layer's stability. Differential advection of moisture and heat is oftentimes needed to produce convectively unstable conditions, and some of the more violent thunderstorms are triggered by release of this form of instability.

Inversions (i.e., frontal, subsidence, radiational) initially have a negative effect on the development of convection by suppressing vertical motion. However, they have the potential for aiding convection at a later time, providing the energy in the form of warmth and moisture trapped beneath the inversion can be released.

Of particular interest is the temperature inversion known as the lid. It is formed by subsidence above a layer of cooler, more moist air. An excellent example occurs when hot, dry (superior) air from the Mexican Plateau is advected across South Texas. The base of the lid is the warmest temperature on the sounding, excluding nocturnal radiation or frontal inversions (Bothwell, 1988). Since the lid is generally near the 700 mb level, the 700 mb temperature should be at or above 10° C. Warmth and moisture build up under the lid during the day, preventing any premature release of energy. The lid can be eliminated through a combination of vertical motion, surface heating, and advection, allowing for the explosive release of energy that can help to fuel severe storms. A second and even more efficient method of releasing energy is called underrunning. This occurs when low-level warm, moist air suddenly emerges from under the edge of the lid into a region where there is no inversion.

It is believed that storm types may depend on a small number of observable parameters that define the storm's environment. Research has consistently pointed to buoyancy and vertical wind shear as the two most important of these parameters, and it suggests they are closely related. Marwitz (1972) found that both vertical wind shear and buoyancy increased in progressing from single cell through multicell to supercell storms. Hodographs of the Alberta hailstorms studied by Chisolm and Renick consistently displayed strong vertical wind shear and high levels of buoyancy while other factors relating to storm growth differed markedly from case to case.

Using a computer model, Weisman and Klemp (1983) examined the effects of increasing vertical shear in storms. Without vertical shear, a cold outflow spreading away from the storm cut off the warm inflow to the updraft and caused the storm to weaken. When vertical shear was increased in low levels (30 kts), the storm formed new cells along the front edge of the cold outflow, with a nearly vertical updraft. Shortly afterward, it also began to weaken, typical of a multicell storm. Increasing the vertical wind shear to 50 kts resulted in the first hint of possible severe weather. The storm split into sustained left-moving and right-moving storms. The right-moving storm contained a cyclonically-rotating updraft in the mid-levels while the left-moving storm had an anticyclonically-rotating updraft. Increasing the vertical shear to 70 kts produced an even stronger updraft. As wind shear was increased to 90 kts, storm splitting still occurred, but the two cells were weaker. In the extreme shear cases, storm development may be stopped altogether.

Tying the two parameters together, the amount of buoyancy needed for a storm to develop increases with increasing wind shear. Conversely, with increasing buoyancy, storms can form at higher shears, with stronger updrafts and downdrafts and more efficient precipitation rates. Thus, strong vertical wind shear is detrimental to weakly buoyant storms while it intensifies more buoyant storms. This optimal development relationship between wind shear and buoyancy is apparently caused by the balance between the downdraft outflow and storm relative inflow. The inflow to the storm must be strong enough to keep the outflow from propagating away from the updraft.

Environments with large buoyancy and weak vertical shear tend to produce storms that build to great altitude, but are short-lived. These could be producers of damaging wind but not likely to develop other forms of severe weather. Environments with weak buoyancy and great vertical shear tend to produce strongly tilted, short-lived convection that does not reach a fully mature state and thus would be unlikely to produce severe weather. Environments with weak buoyancy and weak vertical shear also produce short-lived storms that are unlikely to develop severe weather. It is the environment that has both great buoyancy and great vertical shear that is best prepared to produce severe thunderstorms, especially the supercell.

ANALYSIS OF STABILITY

There are numerous kinds of indices that give a quick estimate of the stability of the air mass. Appendix A explains how these indices are derived and some of the limitations of each. The most widely used, both in operations and in research, is the Lifted Index (there are several versions). It appears to produce the best results and the fewest problems, since it better represents the distribution of moisture in the very important subcloud layer. The forecaster must keep in mind that any index is only intended as a general guide, and must be used with all other available information. Lifted Index values lowering into the -3 to -6 range indicate that the chance of severe weather is increasing. Values between -6 and -12 act as a red flag to alert the forecaster to a day with the potential for activity of extreme severity.

The best advice on determining stability is to plot the entire sounding up to 100 mb. Using the temperature and dewpoints that will be representative of the time of expected maximum heating, adapt the surface and low-level conditions to the location in question. Make any changes in upper-level conditions due to advection and warming as indicated by current and progged charts. Next, using the average mixing ratio in the lowest 100 mb and the expected maximum temperatures for the day, lift the parcel to its LCL and then via a moist adiabat to the top of the sounding. (A Skew-T diagram is used by many forecasters since the size of an area is directly proportional to the kinetic energy that must be supplied to move a parcel along a prescribed path.) Thus, the size of the positive area after lifting the parcel (where the parcel path is warmer than the environment) will indicate how much energy can be made available to a thunderstorm once the instability is released. A large positive area is a hint that severe weather is very possible, if other factors are also favorable.

Determine if a lid is present. If so, its contribution can be evaluated in several ways. Each of the two following methods refers to the attached Skew-T diagram.

Method 1. The ADAP (AFOS Data Analysis Program) program developed by Bothwell (1986) produces, among many other valuable small-scale analyses, a map of the cap (lid) strength over a large area (Figure 3-2). It does so in the following manner. From the warmest point on the sounding (Figure 3-3) that is at least 50 mb above the surface (Point A'), descend via a moist adiabat to the surface (Point A). The temperature at this point is referred to as Θ_{swl}. The surface wet-bulb temperature (Point B) is referred to as Θ_w.

The cap strength is ($\Theta_{swl} - \Theta_w$).

Method 2. Carlson (1983) used two terms, called the stability term and the buoyancy term, to evaluate the contribution of the lid. To calculate the stability term, first determine Θ_{swl} as before. Next, from the highest wet-bulb temperature in the lowest 100 mb (Point C'), descend via a moist adiabat to the surface (Point C). The temperature at this point is referred to as Θ_{wa}.

The stability term is ($\Theta_{swl} - \Theta_{wa}$).

To determine how the lid affects the overall buoyancy of the airmass, first locate the 500 mb wet-bulb temperature (Point D). From this position, descend via a moist adiabat to the surface. The temperature at this point is referred to as Θ_{sw}.

The buoyancy term is ($\Theta_{sw} - \Theta_{wa}$).
The algebraic sum of the two terms is called the Lid Strength Index or LSI. Notice that the stability term gives a value that is generally close to Bothwell's cap strength. A negative (positive) stability term implies the lack (presence) of a lid. A negative (positive) buoyancy term implies air mass instability (stability).

Using either the Bothwell cap strength or the Carlson LSI, determine if the lid can be broken and where this is likely to take place. Remember that underrunning is possible near the edge of the lid. Cap strength and LSI values less than +4 identify areas with a potential for intense thunderstorm development. For the short term, minutes to an hour, these values can even be "advected" by overlaying charts of 850 mb winds. Underrunning is most likely at the edge of the lid in the area where large positive LSI values fall quickly to +3 or less.

As was mentioned earlier, buoyancy is closely related to vertical shear, which can be examined very quickly through the use of a hodograph. A hodograph, as shown in Figure 3-4, is a line connecting the tips of the upper-level wind vectors which are plotted with their origins at the zero point (Doswell, 1989). The line connecting two vectors is the shear between the wind at those two levels. Although small amounts of shear between layers can produce hodograph signatures with loops within loops, it is important to view the broad features of the diagram

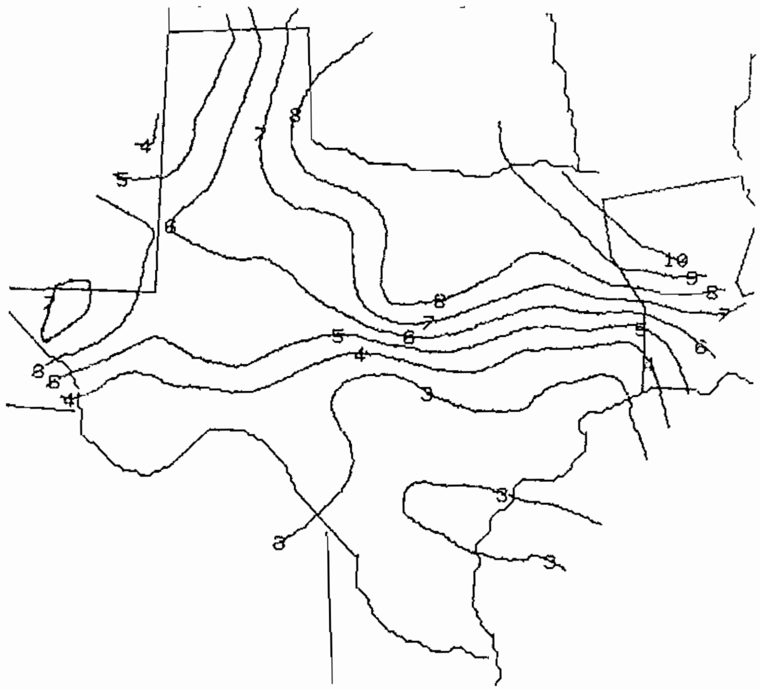


Figure 3-2. AFOS graphic depiction of lid strength (after Bothwell).

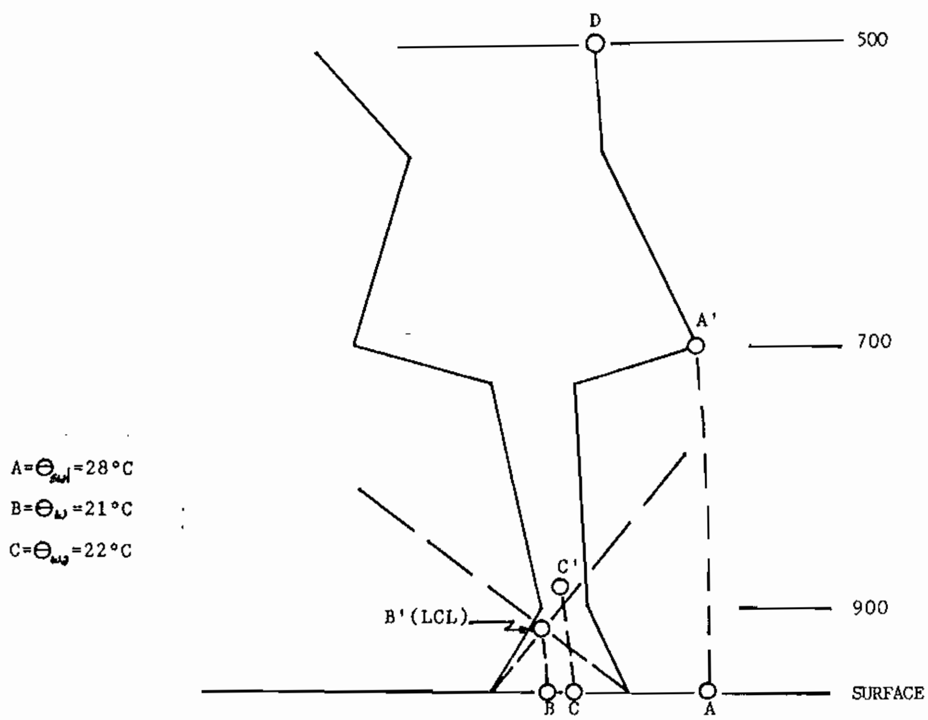


Figure 3-3. How lid strength is computed. Point A' is warmest point on sounding which is lowered moist adiabatically to surface (Point A); Point B is surface wet-bulb temperature; Point C' is the highest wet-bulb temperature in lowest 100 mb which is lowered to the surface (Point C); and Point D is 500 mb wet-bulb temperature (after Bothwell).

and not the small details. Backing of the wind vector with height indicates cold advection and a counter-clockwise turning hodograph. Veering of the wind vector indicates warm advection and a clockwise turning hodograph.

A short, straight-line hodograph indicates weak one-directional shear. This produces short-lived cells whose gust fronts move quickly away from their generally weak updrafts (Weisman and Klemp, 1986). A longer, straight-line hodograph would most likely produce a storm splitting into left-moving and right-moving pairs as mentioned above (Doswell, 1989). Most hodographs, however, are curved, and thus favor one of the split pair. Clockwise turning favors the cyclonic right-mover while counter-clockwise turning favors the anticyclonic left-mover. Weisman and Klemp also noted that a hodograph with a single, relatively small, closed loop was generally associated with a short-lived multicell storm, while the small, open loop was associated with the multicell squall line. With larger loop size (either closed or open), hodographs became increasingly associated with the supercell storm. Rasmussen and Wilhelmson (1983) in their study also found that tornadic storms were associated with large or broad loops, and weaker storms occurred on days with small loops or nondescript hodographs.

One way to relate buoyancy and wind shear to storm strength is through a nondimensional parameter called the Bulk Richardson Number, Ri. This value is displayed as BRN on the AFOS-derived hodograph. The denominator is a measure of the low-level wind shear. The numerator is a measure of potential updraft, downdraft and surface outflow strength. BRN values between 35 and 1100 tend to produce multicell storms while values from 8 to 50 are more likely to produce split storms such as supercells (Weisman et al., 1986).

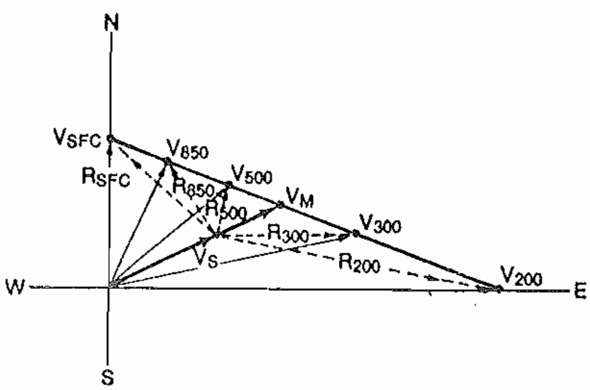


Figure 3-4. Example of a hodograph where Vm is mean wind, Vs is motion of storm, and R represents relative wind at various levels.

CHANGES IN STABILITY

It is important to remember that the following four processes could have changed lapse rates since the time the sounding was made:

1. Non-adiabatic heating and cooling. These sources include radiation, conduction, evaporation and condensation.

2. Advection of air with a different lapse rate. This occurs in the wake of an outflow boundary, or behind a strong cold front.

3. Differential temperature advection due to vertical wind shear. Cooling aloft, combined with warming below, results in decreased stability. In fact, no advection (or even warm advection) in the mid and upper levels can result in decreased stability since the low levels can warm so much more rapidly.

4. Differential moisture advection due to vertical wind shear. An example is the intrusion of dry air in the mid levels. The superposition of dry, more dense air over moist, lighter air results in a destabilization of the air mass. Dry air aloft also provides a method of chilling and accelerating wet downdrafts through evaporative cooling, helping to form the Rear Flank Downdraft.

5. Vertical motion (produced by orographic or convective lift, convergence or divergence). The lifting can occur over a large scale, as with a strong upper-level trough, or on a much smaller scale, such as at the intersection of thunderstorm outflow boundaries. Such lifting can quickly saturate air parcels, producing convection over areas where the moisture field had previously appeared to be very weak.

A parameter distantly related to stability that correlates well with the production of large hail at the surface is the wet-bulb zero (WBZ). When values are below 5,000 ft or above 11,000 ft, large hail occurrence is rare. The optimal values are between 7,000 and 10,500 ft. Higher values indicate the atmosphere is too warm in the mid and upper levels and thus too stable. Also, the hail may melt before reaching the ground. Lower values may mean the atmosphere is too cool in the lower levels and therefore too stable.

MOISTURE

Severe weather generally occurs to the west or northwest of the surface dew point ridge, where dewpoints generally exceed 60° F. The chance of severe weather is further enhanced if a temperature ridge lies to the southwest of the threat area, with the strongest thermal gradient across the threat area. Severe weather is very rare with dew-

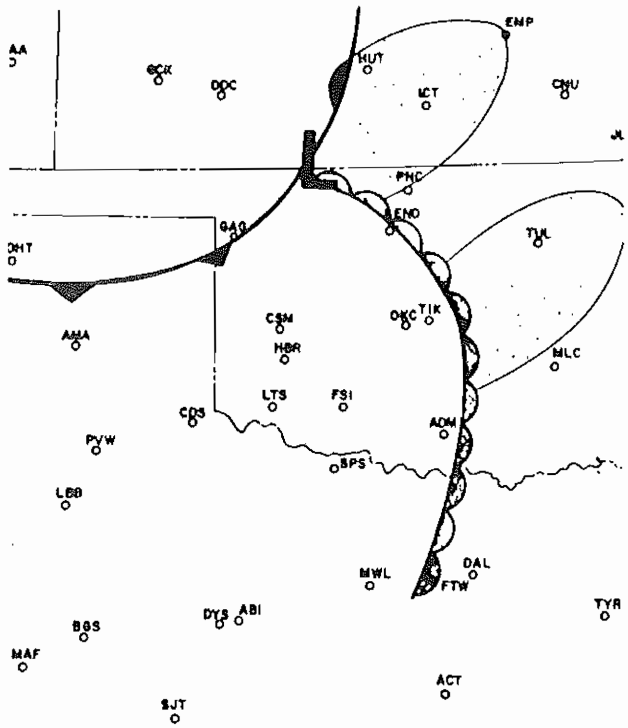


Figure 3-5. Favored locations for the development of severe thunderstorms along the dryline (after Moller, 1980).

points below 55° F. The forecaster must keep in mind, though, that this rule can be broken in unusually strong winter or early spring events. Thunderstorms that occur with overrunning (i.e. warm, moist air lifting over cooler air below) can also produce large hail, but seldom cause high winds or tornadoes.

Severe weather is also more likely to take place to the east or northeast of the axis of highest 850 mb temperatures. The reason for this is that with the hottest and driest air to the west, moisture inflow (and especially moisture convergence) from the east or southeast is maximized. The Low Level Jet (LLJ) is a major source of moisture convergence (fuel) in severe weather outbreaks. LLJ speeds associated with tornadic storms frequently range between 30 and 50 knots. The jet also helps the forecaster to focus more closely on the threat area, since severe weather generally occurs to the left of the LLJ where cyclonic vorticity (the shear term) is greatest. On most days, the LLJ can be located by identifying on a surface chart the axis of highest dewpoints. The LLJ would lie very close to this line.

Another preferred area is just in advance of a bulge along the dryline (Moller, 1980). This bulge (Figure 3-5), where moisture tends to concentrate, usually occurs because dry air in the 850 to 700 mb range is brought to the surface, causing the line to surge forward. The movement of the dryline is quite interesting. During the day, strong heating causes the surface air to mix with drier air in the mid levels of the atmosphere. At the surface, this causes dew points to lower and winds to veer to the west. At night, the mixing ends with the formation of a low-level inversion and surface winds shift back into the east, reintroducing the higher dew points. Thus, the dryline "mixes" to the east during the day and "advects" back to the west during the night. The best way to locate the dryline is to place it where the winds first veer to the southwest and the dewpoints first begin to decrease. This is also an area of very strong moisture convergence. Both patterns are usually accompanied by a weak capping of moisture, heat, and instability in the low or mid levels of the atmosphere.

Moisture convergence can also be enhanced by small areas of rapid pressure falls. Many severe outbreaks have been associated with such patterns. These areas, approaching from or developing to the west of a location, cause low-level wind speeds to increase and surface wind directions to back. The net effect is to bring about an increase in moisture convergence, and focus the convergence into a single point. This is most effective when the size of the pressure rise areas is very small.

ANALYSIS OF MOISTURE

Since severe weather usually occurs to the west of the axis of deepest moisture, in the vicinity of the steepest moisture gradient, it is important to locate both the surface and 850 mb axes of greatest moisture. An isodrosotherm analysis at each level is the best way to define these areas. Next, look at an appropriate sounding to determine the available moisture. For severe weather to occur, the mixing ratio should be equal to or greater than 10 g/kg in the lowest 50 mb.

Mid-level moisture pools can also provide the required fuel for storms. In addition, they may be clues that these areas became moist through condensation provided by lifting. In this way, they could pinpoint areas of strong upward motion as well.

An excellent guide to the inflow of fuel to an area is the moisture flux convergence (chart SMC) and moisture flux convergence change (chart SCC) in the AFOS ADAP program (Figures 3-6 and 3-7).

Trigger Mechanisms

Since there is no way to measure vertical motion directly, it must be inferred from other parameters or features. This can be done by looking for patterns that have been associated with upward motion and convection. Examples of this would be low-level convergence and/or upper-level diffluence. Displays of vertical motion fields determined by the Nested Grid Model (NGM) are available on AFOS.

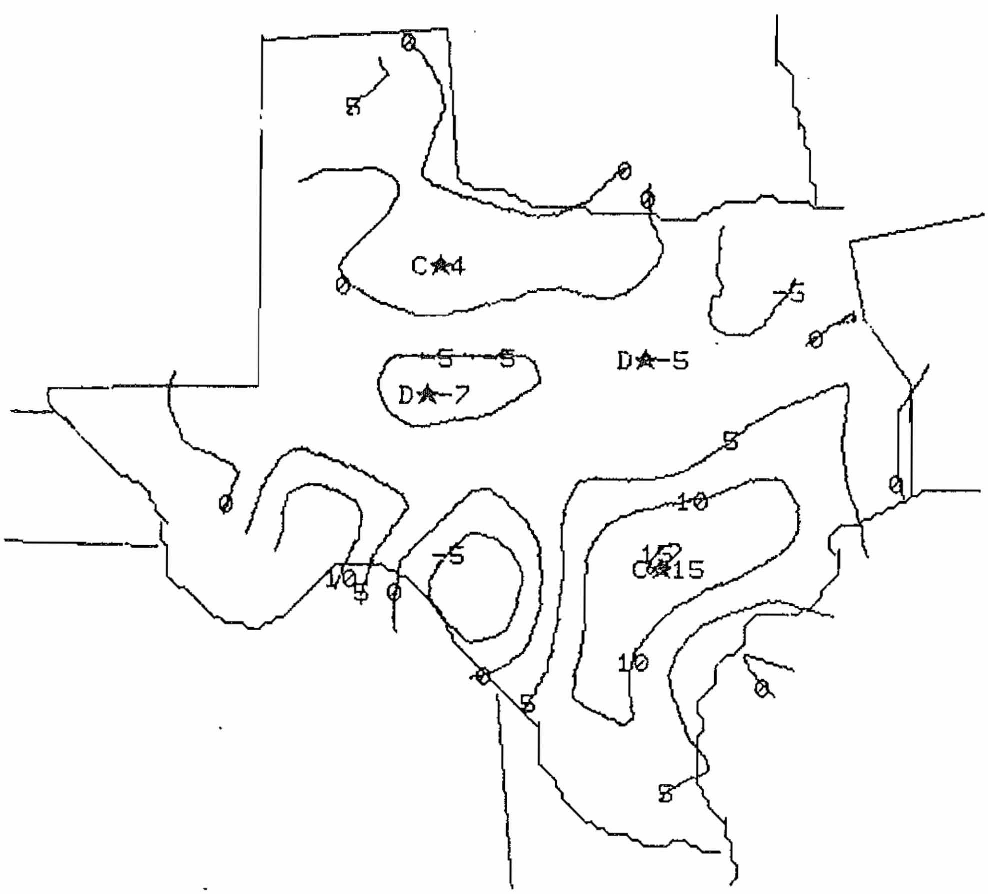


Figure 3-6. ADAP map of surface moisture flux convergence (after Bothwell).

There are a number of classic low-level weather patterns that have been identified with an increased chance of severe weather. As a reminder, patterns that result in convergence near a single point (point convergence; hence point lifting) result in single, isolated severe storms, while producers of convergence along a line (linear convergence and thus linear lifting) will generally result in squall line formation.

Related to low-level pressure patterns, many of the most significant severe weather events take place to the northeast of a deepening sub-synoptic low, as shown in Figure 3-5. Here lifting, moisture convergence and low-level warming are at a maximum, indicating sufficient trigger, fuel and instability for storm development and strong upward motion.

Upper-level systems capable of producing lifting can also be identified at times by areas of cooler temperatures or speed maxes. It is very helpful to have winds of at least 40 to 50 kts in the mid layers (700 mb to 500 mb). Among other reasons, these winds would be higher than the forward speed of the thunderstorms, increasing the possibility for a Rear Flank Downdraft to occur. Directional shear in the lower portions of the mid levels also enhances the chance of the storm to develop a rotational updraft. In the upper levels, winds of at least 50 to 60 knots help to increase the severe potential of a storm, providing the instability is also very strong (Lifted index -5 or lower.) Otherwise, the strong winds would produce a highly tilted, weak updraft. Strong vertical speed shear has several effects on a storm: it rapidly destabilizes the air mass the storm is developing in, it assists in producing divergence aloft (which increases upward motion), and it allows the transport of mid-level momentum to the surface in the form of downbursts. Also, strong winds carry the rain to the front of the storm and away from the updraft, extending the life of the thunderstorm.

Vertical motion can also be inferred using the concept of vorticity. Vorticity is a measure of the cyclonic turning present in the atmosphere. It can be calculated at any level in the atmosphere, either on a constant pressure chart or an isentropic surface. In the past, custom has decreed that the analysis be performed at the 500 mb level, associating Positive Vorticity Advection (PVA) with areas of upward vertical motion. In fact, PVA implies upward motion if and only if 1) PVA increases with height above 500 mb and 2) air parcels are moving faster than the vorticity field. The first requirement guarantees that upward vertical motion will continue above the 500 mb level (absolutely necessary for severe weather). The

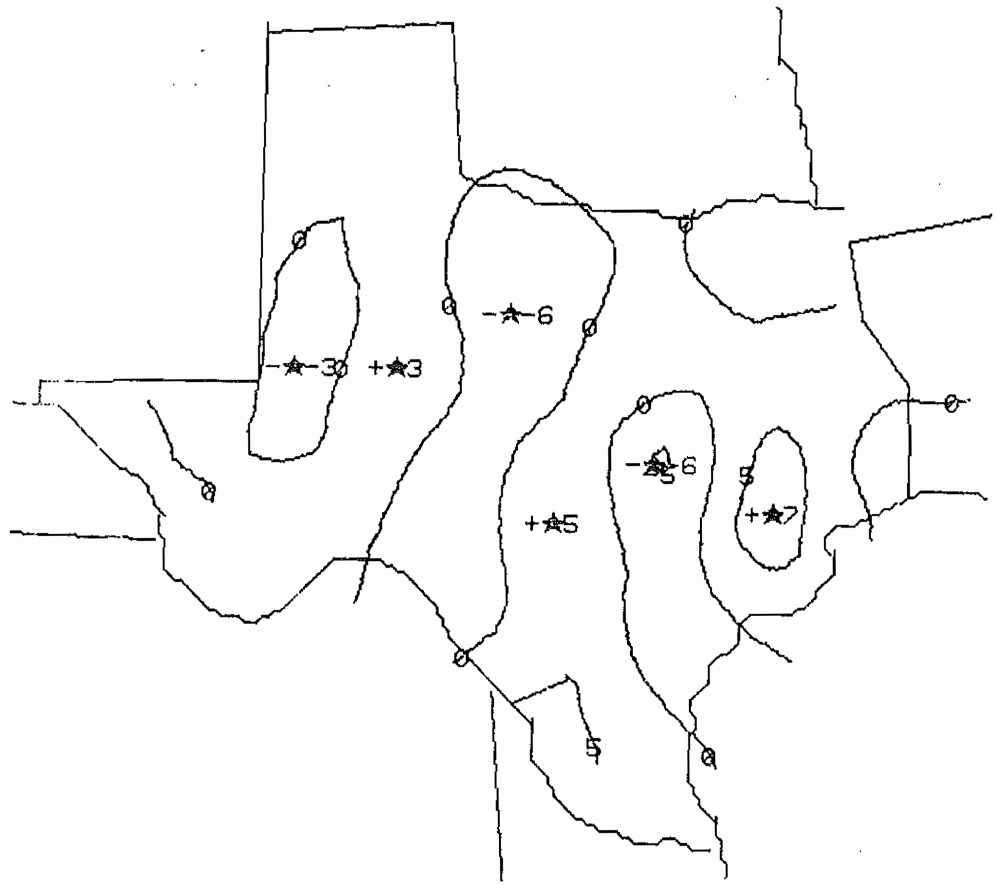


Figure 3-7. ADAP map of surface moisture flux change (after Bothwell).

second requirement explains why fast-moving vorticity maxima may not produce upward vertical motion or convection. It also guarantees that the air parcels will be changing their vorticity to that of the environment as they move through the vorticity field. They do this through divergence. Thus, PVA can lead to convergence in the lower levels and divergence in the upper levels.

Research by Hales (1980) and Maddox et al. (1982b) of several years' worth of severe weather occurrences indicated that, in most cases, PVA was not present at 500 mb. Similarly, they looked into the thermal advection associated with the outbreaks. Instead of seeing cooling at 500 mb, as schooling has generally suggested in the past, either no change or warming actually took place. The most common cause, and best indicator, of upward motion was actually warm advection, both in the lower and mid levels of the atmosphere.

The concept of PVA also suffers from being a 2-dimensional approach. It does not consider the vertical motion of air parcels as they move through the atmosphere, nor does it take warm and cold advection into account. Fortunately, work done by Trenberth in 1978 formed the beginning of a field of study known as quasi-geostrophic theory. He showed that the forcing terms in the omega equation could be grouped and simplified into a single term that involved the advection of vorticity by the thermal wind. Positive advection is referred to as Positive Isothermal Vorticity Advection (PIVA) and negative advection is referred to as Negative Isothermal Vorticity Advection (NIVA). Excellent discussions have been written on this subject by Barnes (1985), Bluestein (1983), Bozart (1985) and Durran and Snellman (1987).

ANALYSIS OF TRIGGER MECHANISMS

Operationally, since the thermal wind in a layer of the atmosphere is proportional to the thickness of that layer, the quasi-geostrophic equation translates into using the thickness field of a layer to advect vorticity in the layer. For best results, the 500 mb vorticity field should be advected with the 700 mb to 300 mb thermal wind, or 700 mb to 300 mb thickness. This is possible using the Quasi-Geostrophic Diagnostics program for micro-computers developed by Mike Foster while at NWS Southern Region Headquarters/SSD. Overlaying chart 5QV (500 mb quasi-geostrophic vorticity) over chart 5QK (700-300 mb quasi-geostrophic thickness) will produce the isothermal vorticity map desired. Since the 700-300 mb thickness chart is not currently available on AFOS, the best depiction of isothermal vorticity is obtained by overlaying the 500 mb vorticity chart on the 1000-500 mb thickness.

On the surface chart, thermal boundaries are areas along which thunderstorms have an improved chance of forming. Locate the cool pockets. These may be areas that have been cloudy during the day or simply the remains of previous thunderstorm outflows. Clearly the intersection of two outflows is a preferred area of upward motion. There is a greater chance of convection as warm winds impinge on the outer edges of these cool bubbles and lift up over them. Thermal ridges often identify the western boundary of severe weather events. Other preferred triggers would include a cold front, a dry line, a sea breeze, or a combination of the these mechanisms.

Locate the thermal ridge on the 850 mb and 700 mb charts. Warm advection, an often neglected producer of low-level upward motion (see Doswell, 1982), is likely to be taking place east of this line.

Small-scale, rapid changes in pressure can be very helpful in identifying areas of increasing severe weather threat. Altimeters are better indicators than sea-level pressure because they do not include a temperature factor and have a much denser spatial distribution. It is best to look for one-hour falls of .03 or more or two-hour falls of .05 or more. Severe outbreaks often follow concentrated areas of rapid falls by as much as 30 minutes to one hour. These are likely areas of strong upward motion. Also, the movement and lowering pressures would also imply increased moisture convergence.
Windspeeds appear to be the best indicators in the highest levels. There are several things to look for. First, the presence of a jet streak can strongly enhance the development of severe storms. It increases the vertical speed shear in the column. This helps to "vent" the air column, just as wind passing over a chimney in the home assists in drawing air up into the chimney. The most favorable areas of the jet for severe weather, relative to the jet core, are the left front quadrant and the right rear quadrant. That is true since divergence is supported in these quadrants, and divergence in the upper levels is directly related to upward motion.

The situation becomes harder to evaluate in the case of a strongly curved jet. If the jet is cyclonically curved, strong divergence still occurs in the left front quadrant, with strong convergence in the right front quadrant. No clear statement can be made about either of the rear quadrants. Conversely, if the jet is anti-cyclonically curved, strong divergence occurs in the right rear quadrant with strong convergence in the left rear quadrant, and no clear statement can be made about either of the forward quadrants.

The interaction of two jets can be complicated. When two polar jets approach each other, the effects of the juxtaposed quadrants could either be additive or subtractive. Only if the effects are additive can a clear statement be made. In another example, lifting can be greatly enhanced if the divergent quadrant of an upper-level jet is superimposed over the convergent quadrant of a low-level jet (Ucellini and Johnson, 1979). When a southward-moving polar jet approaches a subtropical jet, experience has shown that severe weather will generally take place to the north of the subtropical jet.

Convergence in the right front quadrant of a subtropical jet enhances very strong subsidence that results in a near-dry adiabatic lapse rate and a strong mid-level capping lid. In fact, this is a potentially explosive situation (with a very large positive area evident on the sounding), providing sufficient moisture is available and a strong enough lifting mechanism is present to remove the cap and release the trapped energy. McGinley (1986) also notes that this right front quadrant of a jet maximum is a preferred region for the generation of gravity waves. These fast-moving waves (with forward speeds on the order of 20-120 kts) are quite difficult to track on our current network of conventional observations and oftentimes just as difficult to detect on satellite. Nevertheless, they are thought to be possible initiators of convection, but with associated vertical motions only strong enough to break weak caps at best.

Chapter 4

EVENT OPERATIONS

This section deals with the latest research and guidance on how to detect and identify the weather parameters associated with severe thunderstorms. All of the pre-event analysis of upper level charts, soundings and all background information on stability, moisture and potential triggers should have already been completed. This background information will have provided the first clues as to whether the potential exists for severe weather, as well as some hints as to the location, timing and type of threat. This chapter deals with the final pieces of information needed before watches, statements or warnings are issued.

There are four general sources of these final details. They are radar observations, satellite imagery, mesoanalysis and spotter reports. The most conscientious forecaster will call on every available clue from the sources and put them together into a unified picture of what is taking place in the atmosphere.

Radar Information

As we have already discussed, thunderstorms move through an evolutionary process with some developing further than others as they produce severe weather. During the later part of the 1970's, laboratory research and field observations were combined in an attempt to learn more about this process and the final product known as the supercell.

The key to the potential of a storm to produce severe weather was found in the strength of its updraft. Although no direct value of updraft speed or strength could be obtained by conventional radar, a general indication could be inferred by the storm structure and changes to storm structure taking place within the storm itself. Les Lemon while at National Severe Storms Laboratory devised a method by which radar could be used to monitor storm structure changes in thunderstorms as they progressed through the evolutionary process. He noted that as storms developed from multicellular non-severe to multicellular severe they began to exhibit predictable changes in their internal and external structures.

For the purpose of clarity as regards Lemon's work, the term "low-level" will refer to the layer from the surface to 5,000 feet and the term "mid-level" will refer to the layer from 16,000 to 39,000 feet.

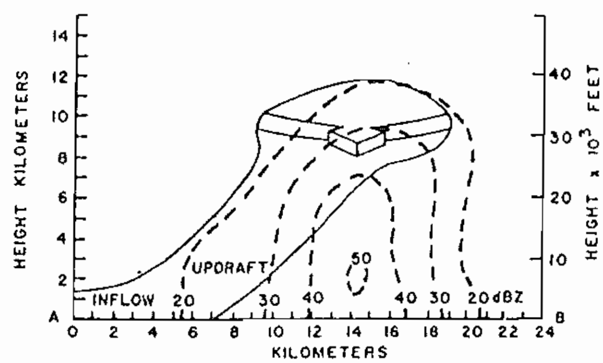


Figure 4-1. The multicellular, non-severe storm has a weak updraft that leans slightly forward (Lemon, 1980).

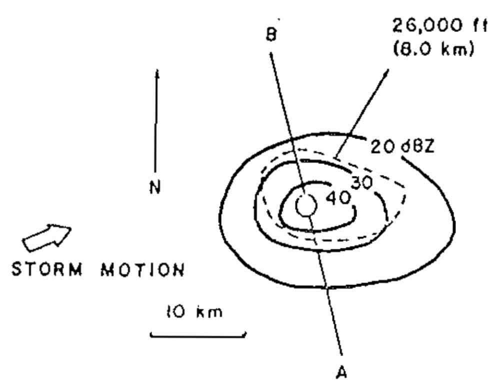


Figure 4-2. The storm top is over the most intense low and mid-level reflectivity areas (Lemon, 1980).

The weak-to-moderate updraft associated with the multicellular non-severe storm is forced by the stronger winds aloft to "lean" slightly forward, or toward the downwind side, as shown in Figure 4-1. This storm will have its highest top over the most intense mid-level and low-level reflectivity areas (Figure 4-2). It will not exhibit any strong gradients of reflectivity in its low-level echoes.

The updraft in a multicellular storm strengthens as the storm becomes severe. The storm develops a strong reflectivity gradient along the low-level inflow side (Figure 4-4). This is simply because the stronger updraft (Figure 4-3) requires a more intense inflow into the storm. The storm's updraft will change from forward-leaning to a nearly vertical updraft. The fact that the updraft is more intense and more vertical and the fact that it is anchored to the inflow side of the storm means that all building now takes place over the low-level inflow. Because of this, the storm top will shift from over the center of the storm to over the strongest low-level reflectivity gradient. The mid-level echo will also shift over the low-level gradient. This mid-level echo will begin to project out beyond the low level reflectivity gradient, usually by 3 to 14 nm. Thus, an echo-free area will become identifiable under this mid-level echo. This is the updraft of the storm and is known as the Weak Echo Region (or WER.) As they become severe, most multicellular storms will also begin to slow their movement, and may even make a turn toward the right or left.

Figure 4-5 shows how, as the updraft intensifies further, the storm reaches the supercell phase. Despite generally

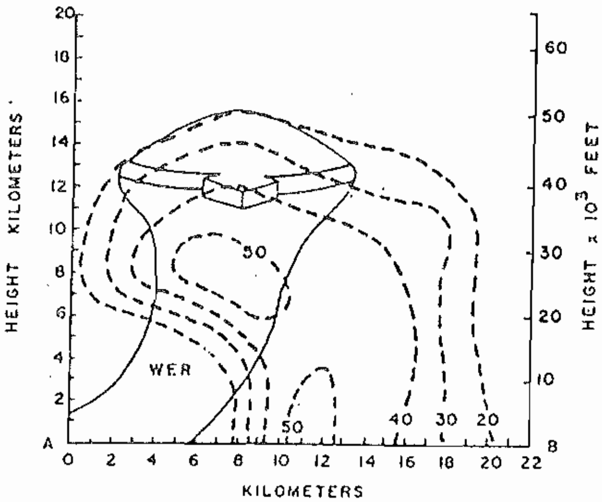


Figure 4-3. As the multicellular storm becomes severe, its updraft intensifies and becomes more vertical (Lemon, 1980).

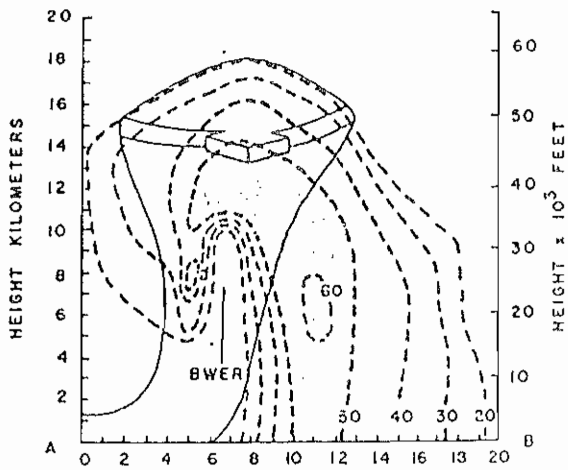


Figure 4-5. Moisture is lifted too rapidly by the intense updraft for condensation to occur. This forms a (bounded) weak echo region, or BWER, in the mid-level echo (Lemon 1980).

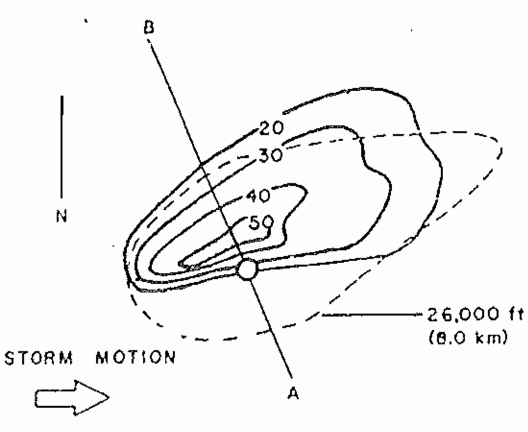


Figure 4-4. The strengthened updraft concentrates moisture along the storm's inflow side, producing a strong reflectivity gradient (Lemon, 1980).

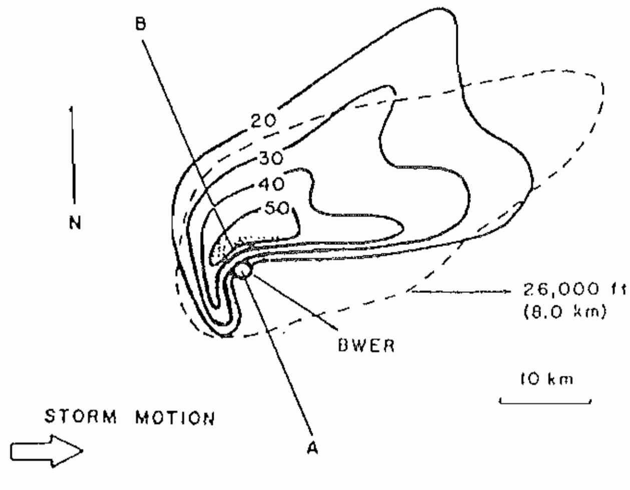


Figure 4-6. The mid-level echo spreads out from 3 to 14 nm beyond the low-level echo. The BWER is located in the center of the storm updraft (Lemon, 1980).

strong environmental winds, the updraft rises vertically over the low-level inflow. The WER, the storm's strong updraft, may be strong enough (with upward velocities high enough) as it moves up through the mid-level to form an empty hole in the radar echo. This is because the updraft has too little moisture or debris in it to produce a radar return, or may be because the upward speeds of parcels are too great for condensation to occur. This "hole" feature is known as the Bounded Weak Echo Region, or BWER (Figure 4-6). It may not be detectable if the radar is not close enough (usually within 60 nm).

Two events are often observed as the supercell begins its tornado production. The first is a lowering of the thunderstorm top. This occurs because the RFD has struck the surface and wrapped around the inflow to the updraft. As

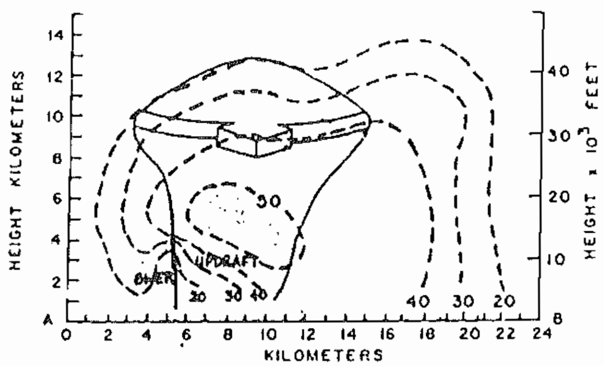


Figure 4-7. As the inflow to the updraft is cut off, the updraft weakens and the top often lowers (after Lemon, 1980).

the fuel to the storm is slowly shut off, the updraft weakens and the storm top lowers (Figure 4-7). The lowering of the top can be mistakenly interpreted as a weakening of the storm, when in fact, it signals the most dangerous time in the life of the supercell.

A second event often associated with tornado production is also produced by the wrapping around of the RFD. This can appear as a concavity at the rear of the thunderstorm with a pendant echo wrapping around the concavity. The storm top will generally be located above this pendant. This is the celebrated "hook echo" which in many cases is associated with the production of tornadoes.

Then, to become severe, two things must occur, and they must occur nearly simultaneously. The storm top must shift from over the center of the low level echo to over the reflectivity gradient on the inflow side. Also, a well-defined mid-level echo must develop over the low-level reflectivity gradient. Any storm that meets these two criteria is capable of producing severe weather.

SEVERE THUNDERSTORM EVOLUTION

The general evolution to a severe thunderstorm is as follows:

A mid-level echo increases rapidly in size and/or intensity.

A WER starts to develop.

Large hail is usually observed within 30 minutes of WER formation.

The echo top shifts from over the storm center to over the low-level reflectivity gradient on the inflow side (and over the WER.).

The low-level echo core shifts toward the inflow and updraft side.

If the updraft is intense enough and the radar close enough (within 60 nm), a BWER may be detected.

The largest hail is usually observed at this time.

The right rear flank often swings southward, forming a pendant.

The echo top reaches its greatest vertical extent.

The RFD is now swinging forward, wrapping around the inflow to the storm updraft (BWER).

Deprived of its inflow, the BWER now begins its collapse.

The mid-level echo surrounding the pendant begins to lower and increase its reflectivity (and often decreases in size.)

The low-level pendant completes its wrap-up and disappears.

The low-level echo often increases in size and weakens.

The echo top lowers by 5,000 to 25,000 feet.

The time between the wrap-up of the pendant and the lowering of storm tops is the most likely time for tornado production.

The echo top shifts back to almost over the low-level echo core.

The storm is quite likely no longer severe.

New updrafts may form and begin the building cycle again.

The goal of the radar analyst is to interrogate echoes as quickly and efficiently as possible to determine whether severe weather is indicated. To accomplish this goal, the operator needs a technique that will reveal pertinent structural data in a minimal amount of time. Mr. Lemon has developed such a technique. The following outlines this technique along with a faster one used by the National Weather Service. Both use horizontal and vertical scans to check for a shift of the echo top and development of a mid-level overhang as indicated above.

LEMON'S TECHNIQUE

Lemon's basic technique
1. Complete a low-level scan.
   Outline the low-level contour of the cell.
   Look for strong low-level reflectivity gradient on the inflow side.
2. Complete a mid-level scan (12,000 to 18,000 feet.)
   Determine if a mid-level echo (of at least VIP 4) exists overhanging the low-level reflectivity maximum.
3. Complete a high level scan near the storm top.
   Mark the storm top.
4. Take care to minimize the time interval between echo tracings (due to rapidly moving storms.)

WRIST TECHNIQUE

The National Weather Service WRIST Technique (Weather Radar Identification of Severe Thunderstorms) permits rapid interrogation of radar echoes. Each cell can be interrogated in 60 to 90 seconds, and many non-severe cells can be weeded out in the first 5 seconds.

1. Mark top of chosen cell with a dot.
2. Compare location with surface echo. If the storm top is over the inflow side or over a strong relectivity gradient along that side, continue. Otherwise, terminate.
3. With antenna rotating at 3 rpm, trace the VIP 2 contour on the inflow side.
4. Draw a dashed line parallel to the inflow side 3 nm out from the VIP 2 contour.
5. Position the antenna so the sweep intersects the 3 nm line in part of the cell most likely to possess a mid-level (16,000 to 39,000 feet) overhang. Terminate if no mid level overhang is detected.
6. With antenna rotating at 3 rpm, interrogate area of maximum overhang to determine if VIP level 4 or greater overhang exists within the 3 nm extension. If such exists, issue a warning.

Here are some other clues on using radar to identify potentially severe storms.

Severe weather is most likely to be associated with isolated storms. This is because these storms become so intense that they draw moisture (and upward energy for vertical motion) away from surrounding echoes. The nearby echoes usually begin to fall apart and dissipate, leaving the severe storms by themselves. This is common, even with lines of echoes. Within the line, the cell that produces severe weather will develop such a strong updraft that all low-level inflow for several miles will be directed into it. There will usually be no other storms that can form on its inflow side. This helps to explain the following rules.

RULES ASSOCIATED WITH LINES

Severe weather is most likely to occur with:

a. isolated cells ahead of lines,

b. cells at a bulge in a line (crest of a Line Echo Wave Pattern- LEWP),

c. cells on the inflow side (usually south side) of a line,

d. cells north of a break in a line (assuming inflow side is south), and

e. cells that move fastest in a line.

Tornadoes rarely occur in lines of thunderstorms. Those that do are generally small, weaker and short-lived.

RULES ASSOCIATED WITH CELLS

Severe weather is most likely to occur with:

a. cells that have changed movement significantly—including cells that exhibit direction changes (right/left movers) and cells that exhibit speed changes (increase or decrease),

b. cells that exhibit a "V notch" in the low-level downwind echo (flow aloft is deflected on either side of intense blocking updraft, carrying precipitation downstream along two distinct bands),

c. cells with VIP 5 levels above 30,000 feet,

d. splitting echoes with separate WER's that deviate to right and left (the right mover is usually slower moving and more likely to be severe), and

e. supercells as opposed to multicellular storms.

Research further suggests that 25 percent of multicell severe storms produce large hail while 80 percent of supercells produce large hail.

No clear statement can be made about the effects of a merger of two cells. The resulting updraft can be enhanced—if the inflow areas are brought together. However, if the outflow of one undercuts the inflow of the other, the result may be the death of one or both of the cells. More often than not, the merger of two cells results in no well-defined change in echoes.

Satellite Information

Satellite imagery is another source of latest detailed information about the nature of thunderstorms. It is limited, however, in its usefulness. Because severe thunderstorms generally produce a great amount of cirrus, it is difficult (if not impossible) to obtain information on these storms, other than about their attendant high clouds.

One formation that has been very reliable in indicating the possibility of severe weather is that of the Enhanced-V. This feature is a V-shaped cold area in the top of the storm, with the V portion open toward downstream. The V is best located using the MB curve on IR imagery. It is thought that this formation is due to a colder overshooting top. Downward motion that would take place in the lee of such an overshooting top would likely produce subsidence. This, in turn, creates the "warm" spot (Figure 4-8) in crest of the V. The Enhanced-V is generally associated with large hail.

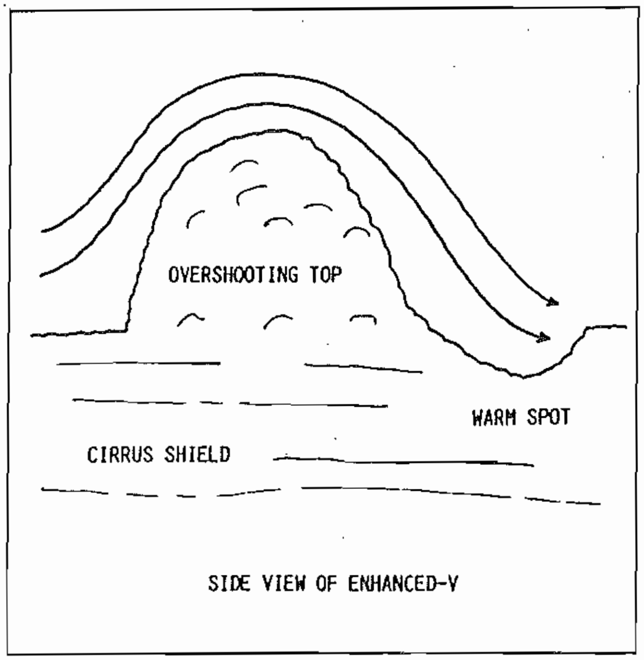


Figure 4-8. The combination of a cold overshooting top and a warm spot produced in the lee of the cold top produces the Enhanced-V signature (after McCann, 1984).

A four-month study was undertaken, using this idea as a warning tool. Of the Enhanced-V formations observed, 70% were associated with severe weather. Furthermore, using this as an indicator produced both median and mean lead times of 30 minutes. This feature is not present in all severe thunderstorms, however. During the study, only about 25 percent of the severe storms exhibited this feature. Thus, with a Probability of Detection of 25% and a False Alarm Rate of only 30%, the Enhanced-V is a sufficient but not necessary condition for severe weather.

There are some other limitations on its use. First, it must be associated with a growing thunderstorm. That is, the colder anvil parts of the storm must be expanding, or the storm's highest top must be cooling. The positioning of the location of severe weather is also somewhat of a problem. Because of parallax, the following procedure is recommended. First, locate the apparent position of the storm's highest top. The warning should be issued for a position between 3 and 15 miles to the south or southeast of this location. Finally, the storm should still be considered severe as long as it is still growing, even if the V figure disappears.

Collapsing tops are also possible indicators of severe weather. Remembering back to the discussions of the evolution of multicellular severe and supercell thunderstorms, rapidly decreasing tops are often associated with production of severe hail or wind. This could even signal the development of a tornado. Overshooting tops are also indicators of storms with great potential for severe weather. The top of the cirrus shield is generally very flat, because it is capped by the tropopause, a very stable layer in the atmosphere. The presence of an overshooting top indicates that the updraft in the associated storm was so intense that it forced cloud parcels to penetrate this stable layer, and probably did so by several thousand feet to have been visible on satellite.

Storms with flanking lines are another good indicator of a potentially severe thunderstorm. This flanking line is an area where low level winds are converging with the resulting warm, moist air being drawn into the rear portion of the thunderstorm. The fact that such amount of inflow can be gathered by the storm testifies to the presence of a very strong updraft. Clearly this storm would be ingesting fuel at a rapid rate.

Analysis Information

The third source of information on potentially severe weather is found in analysis. It is understood that the immediate (or nearby) threat of severe weather generally robs the forecaster of the much-needed time to do additional analysis. However, quick plots of the most favorable areas and analysis that is confined to the most promising parameters will not require as much time as the more formal analyses.

Frequent surface charts are the first suggestion. These are used to define areas each hour or two where severe weather is most likely to take place. From the previous discussions, some of the suggestions would be to keep a close watch along and west of the surface moisture ridge, near moisture pockets and along and east of the surface thermal ridge.

The locations and movements of pressure boundaries are important. This would include sub-synoptic lows, surface dry lines (especially with bulges), fronts and thunderstorm outflow boundaries. Two-hour altimeter changes are often very helpful in zeroing in on locations where cluster or corridor type severe events may later take place. Look for continuity of the pressure-fall centers. Some will be short-lived while others can last for several hours. The possibility that a pressure-fall center could move to intersect a density boundary, such as a thunderstorm outflow boundary or line of storms greatly enhances the chance of thunderstorm development. In general, the smaller and more compact the pressure-fall center, the more severe will be the convection. Also, the location and movement of density boundaries, such as the cool remains of earlier convection or pools of moisture can indicate preferred areas of severe thunderstorm development.

One parameter that can be calculated by personal computer is that of moisture convergence. This has been very useful in determining areas where strong vertical motion is, or is not, likely to take place. The term "moisture convergence" (or sometimes, "moisture flux convergence") is made up of two terms. The first is wind convergence multiplied by the mixing ratio. The second is moisture advection. This is essentially a measure of how rapidly warm, moist air (thunderstorm fuel) is being drawn into a small area. Convection is more likely to occur when the values are higher, and also when the values are increasing hour by hour. There is no way of obtaining a single value that would indicate precisely when convection would begin. The first parameter, the product of wind convergence and mixing ratio, is usually the greater of the two.

Storm Spotter Information

The final source of real-time clues on severe weather comes from spotter operations. The forecaster must be familiar with the terms that will be used by spotters, as well as the limitations and problems that spotters may face. Since heavy rain, hail and damaging winds can be identified by the public, spotters are most likely to be called into action when the severe weather threat is that of tornadoes. For this reason, we will concentrate on the supercell thunderstorm.

The supercell thunderstorm depicted in Figure 4-9 by Doswell is moving from left to right. The updraft is nearly vertical, so if you were to draw a line downward from just in front of the highest top, it would effectively separate the front (outflow) portion of the storm from the rear (intake) portion. As a supercell thunderstorm approached a spotter, the first effect the spotter would feel would likely be a warm wind blowing from his location toward the storm. This is fueling the storm and can be felt while the storm is still several miles away. A few minutes later, as the storm drew closer, the spotter would likely detect the colder outflow coming toward him from the storm. This is the lead edge of gusty winds, also known as the straight-line winds or gust front. It could be strong enough to be severe or cause damage.

As the storm neared, the sky would look quite dark and a low sloping cloud deck (the shelf cloud) would appear first, likely accompanied by light rain, then heavier rain as the storm moved over the spotter. This low cloud can become detached from the storm, preceding it as a low rolling deck parallel to the ground. This phenomenon is usually associated with very strong winds at the surface and the cloud is referred to as a roll cloud. This is rotating about a horizontal axis, not a vertical axis as a tornado rotates. This cloud can be confused at times because it rolls. It is only a hint that strong straight-line winds could follow.

As the storm continued over the spotter, the rain would generally get heavier. The low clouds above the spotter would be ragged-looking, indicating the cold outflow taking place in the lower front of the storm. Behind the area of heavy rain is the preferred location for small hail, and later large hail. The supercell does not always produce hail, but if it does, it usually occurs behind the heavy rain area. By this time, the forward portion of the storm would have moved past the spotter and he would be in an area of light winds, between the outflow in the front of the storm and the inflow in the rear of the storm. In fact, many experienced spotters have noted how calm conditions were in this part of the storm.

Then, suddenly the calm has ended. The spotter would now see above him a dark smooth base to the thunderstorm. This indicates that warm air is being drawn into the storm in this area. It is referred to as a rain-free base, and serves as the main intake for the storm's fuel. This base often extends for several hundred yards to even a mile or more, with numerous smaller cells building in a flanking line above it. These flanking line storms indicate that low-level convergence is taking place along the base. They continually feed on this energy as they propagate up the line and disappear into the main storm tower.

In the case of a very intense updraft, the warm moist air being drawn into this rain-free base often condenses as it lifts. It then forms a lowering of the base called a wall cloud. This wall cloud is connected to the base of the thunderstorm in the inflow area and may be as large as 2 or 3 miles wide. The presence of a wall cloud indicates that the associated storm has a strong updraft and a high potential for producing severe weather.

If a "valid" wall cloud suddenly begins to rotate, a tornado warning should be seriously considered. To be valid, the wall cloud must persist for several tens of minutes, be connected to the rain-free base and be located on the low-level inflow side of the storm. It must also have a surface based inflow (there must be warm air being drawn into it, as indicated by a smooth underside in at least part of the wall cloud.) Rotation usually occurs as the Rear Flank Downdraft strikes the surface. It can be seen as a clearing slot that begins to appear to the west of the flanking line, trying to wrap around the wall cloud intake area.

Sometimes, an extension of a wall cloud can be seen pointing toward the forward rain area. This is called a tail cloud. It does not rotate, and should not be confused with a tornado or funnel. The second law of the spotter is "If it don't spin, don't call it in!!" The first law of the spotter is to have a " 'fraidy hole" handy.

There are many problems that make the job of spotting even more difficult. One of the worst is that trees,

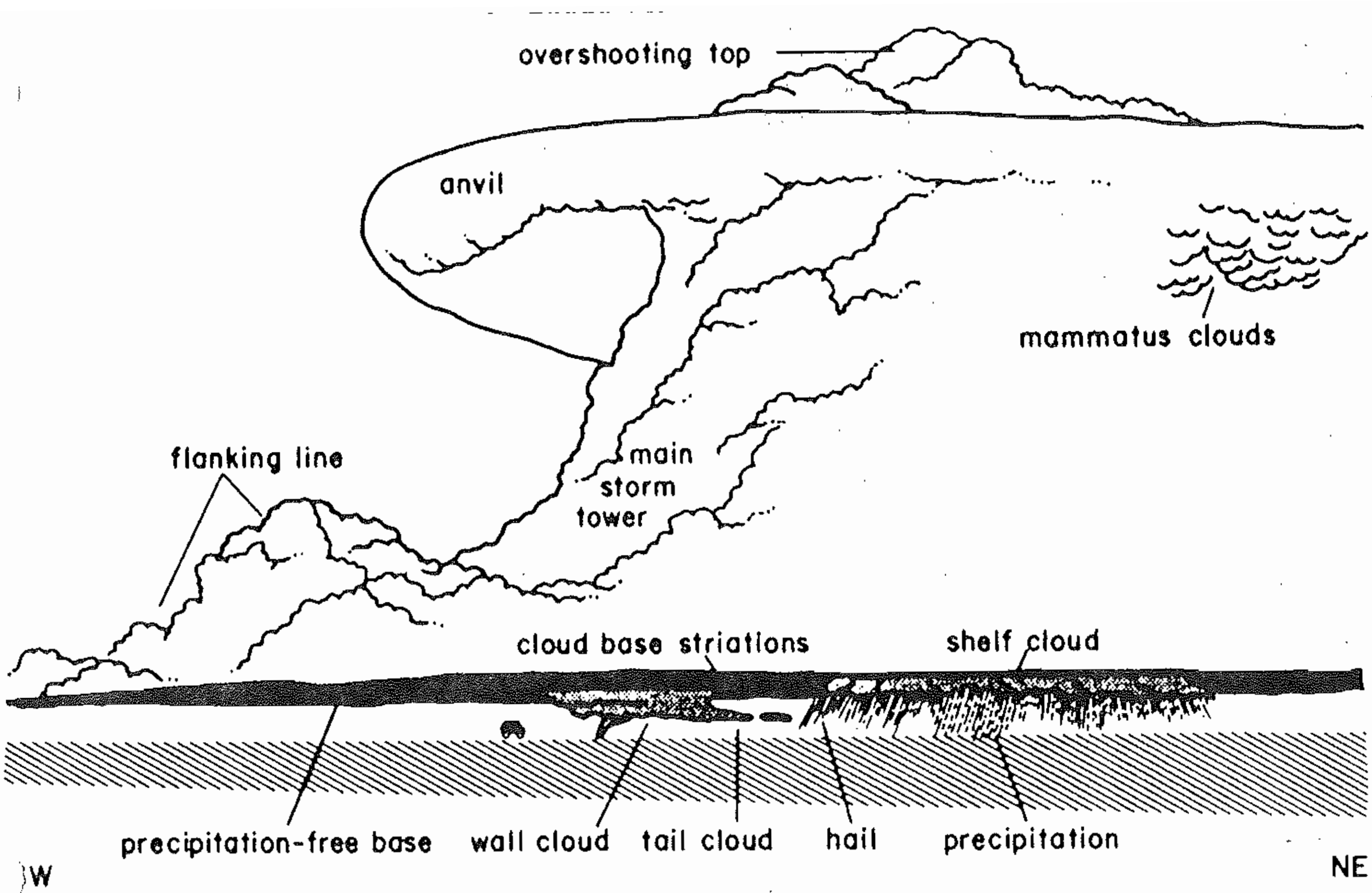


Figure 4-9. Side view of a supercell (after Doswell and Birnam). Spotters would note the outflow wind/rain/hail in the front of the storm and rain-free base and wall cloud in the storm's rear. The flanking line, rear-anvil overhang, and overshooting top also mark this storm as having potential for severe weather.

buildings, or even areas of rain can get in the way of the spotter. In fact, a small tornado can be almost completely hidden in an area of moderate rain. Tornadoes that occur at night are especially hard to detect. Spotters have learned to take advantage of lightning flashes, and to key in on flashes of light from power lines that the tornadoes have knocked over.

Tornadoes sometimes fail to obey the orderly rules we have established. They can occur in the wrong part of the storm, such as the side or in the front beneath a shelf cloud (on the storm outflow.) These tornadoes are minimal tornadoes and are not the threat to life that the more intense tornadoes are. The most likely part of the storm for a significant tornado to be seen is under the rain-free base, usually in the area of the wall cloud, if one exists. Tornadoes also occur in multiples. Multiple vortex tornadoes have been observed where one or more small tornadoes rotate about a larger tornado. It is also possible to have two large tornadoes rotate around each other.

Other spotting problems arise from seeing cloud features that are misleading. Without proper training, such cloud features as roll clouds and tail clouds could be mistaken for tornadoes. Low broken clouds (stratus fractus) can be mistaken for wall clouds. (Remember, the cloud feature must be attached to the thunderstorm to be a wall cloud or tornado.) The greatest problem, though, is due to rain or hail shafts. Dense shafts, when viewed from a distance, can be seen to descend from the base of the thunderstorm, just as a tornado would do. The only way to be sure is for the spotter to continue to watch the feature until he is sure whether it is rotating or not. Also, rain and hail shafts generally do not hold together well and last only a very short time.

With these terms and ideas in mind, the forecaster is prepared to receive and evaluate the information received from spotters. There are some specific features that can clue the forecaster that storms are intensifying, becoming severe or perhaps even preparing to produce a tornado.

There are several visual signs that a storm is intensifying. Recalling that the strength of the updraft is the best clue to the intensity of the storm, it is important when one updraft becomes dominate in a multicell storm. A storm that develops a rain-free base in its rear quadrant has also begun to intensify, just as one that forms a well-defined flanking line. Another sign of intensification is the development of an anvil overhang spreading toward the rear of the storm. This is being produced in the direction opposite to the strong upper level winds, indicating a very strong updraft. This overhang should have solid, well-defined edges to be meaningful. Finally, the presence of a well-defined shelf or roll cloud can indicate the storm has produced very strong outflow, and possibly straight-line winds.

One of the visual signs that a storm has become severe is the development of an overshooting top. Again, this indicates an intense updraft has forced cloud material into the very stable tropopause layer that caps the storm cirrus. Any storm that abruptly changes its movement, (turns to the left or right or slows up or accelerates) is no longer responding to environmental winds. This often means the updraft has reached the critical rate of rotation that will result in the development of a Rear Flank Downdraft and damaging winds. The Rear Flank Downdraft appears as a clearing arc that surges forward behind the rain-free base.

The formation of a wall cloud below the rain-free base is a sign that a storm has potential to produce severe weather. Wall clouds are generally 1 to 3 miles wide, and connected to the thunderstorm, generally under the rain-free base. To be a "valid" wall cloud, it must last for tens of minutes, be located on the low-level inflow side of the storm, and have a surface-based inflow. This means that warm air is flowing into the wall cloud, as evidenced by smooth bases in at least a part of the wall cloud. As often seen in the forward outflow of a storm, ragged bases indicate cold outflow.

The most reliable warning that a tornado is about to occur is the presence of a rotating wall cloud. Again, this must meet the criteria for a "valid" wall cloud as mentioned in the paragraph above. A rotating wall cloud can precede the associated tornado by 20 to 30 minutes.
These are the tools offered by the spotters, or by looking out the window of the weather station in some cases.

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Appendix A
A REVIEW OF STABILITY INDICES

The most commonly used stability indices are described below.

Showalter Index
Using the 850 mb temperature and dew point, find the LCL. From the LCL lift the parcel along a moist adiabat to 500 mb. The parcel temperature at this level is subtracted from the environmental temperature.

The Showalter Index gives poor results when the 850 mb moisture is not representative of available low level moisture, when a strong inversion is present to cap the lifting or when rapid temperature changes occur.

Lifted Index
Determine an LCL from the expected maximum temperature and average surface to 3,000 feet mixing ratio. From this LCL, lift the parcel via a moist adiabat to 500 mb. Subtract its temperature from the environmental 500 mb temperature.

The Lifted Index, generally more negative than the Showalter Index, does not consider inversions and temperature changes, but performs better than the Showalter Index.

Best Lifted Index
This is computed just like the LiftedIndex except that instead of surface to 3,000 feet as the moist layer, several different values are tested and the one with the lowest index is selected.

The Best Lifted Index is an improved version of the Lifted Index.

K Index
Subtract the environmental 500 mb temperature from the 850 mb temperature. Add the 850 mb dew point. Subtract the H7 dew point depression.

The K Index considers moisture in the low and mid layers. It is most useful in pinpointing the threat of heavy rainfall due to a moisture-rich environment.

Vertical, Cross and Total Totals Indices
Vertical Total — 850 mb temperature minus 500 mb temperature
Cross Total — 850 mb dew point temperature minus 500 mb temperature
Total Totals — Add the Vertical and Cross totals algebraically.

Once again, 850 mb moisture may be representative of low level moisture. This index also fails to consider strong inversions or temperature changes.

SWEAT Index (Severe Weather Threat)
SWEAT = 12*D + 20 * (T - 49) + f8 + f5 125*(S + 0.2)
where
D = 850 mb dew point (°C)
T = Totals Totals index
f8 = 850 mb wind speed (kts)
f5 = 500 mb wind speed (kts)
S = sin(500 mb wind direction - 850 mb wind direction)
Omit S if < 0 or 500 mb wind direction not between 310 and 310
or 850 mb wind direction not between 130 and 250 degrees
Assume that D > 0 and T > 49.

SPOT Index (Severe Weather Potential)

SPOT = A + T + Td + W - 110

where

T = surface temperature (°F)

Td = surface dew point temperature (°F)

A = (30.00 - altimeter setting) * 100 if altimeter ≥ 29.50

A = (30.00 - altimeter setting) * 50 if altimeter < 29.50

W is read from the following chart, where V = surface wind speed

	Surface wind direction					
	001-069	070-140	141-200	201-230	231-260	261-360
60 ≤ Td	0	V	V if tstm ocrg 2* V if no tstm	V	V	-2 * V
60>Td>49	0	V	V if tstm ocrg 2* V if no tstm	.5 * V	-1 * V	-2 * V
Td ≤ 49	0	V	V	0	-2 * V	-2 * V

The SPOT Index was developed empirically, using severe weather outbreaks by the Air Force team of Miller and Maddox. It can sometimes pinpoint a threat several hours in advance (Wichita Falls Tornado, 1979). Moisture discontinuities (dry lines or deep marine layers along a coast) can cause problems with this index.

Energy Indices

EI1 — The parcel in the lower 150 mb of the sounding with greatest wet bulb potential temperature is raised to 400 mb while entraining environmental air at a rate that provides a 60 percent increase in mass over a 500 mb ascent. The index is the net energy area (positive minus negative).

EI2 — Same as EI1 except parcel is lifted to the Equilibrium level.

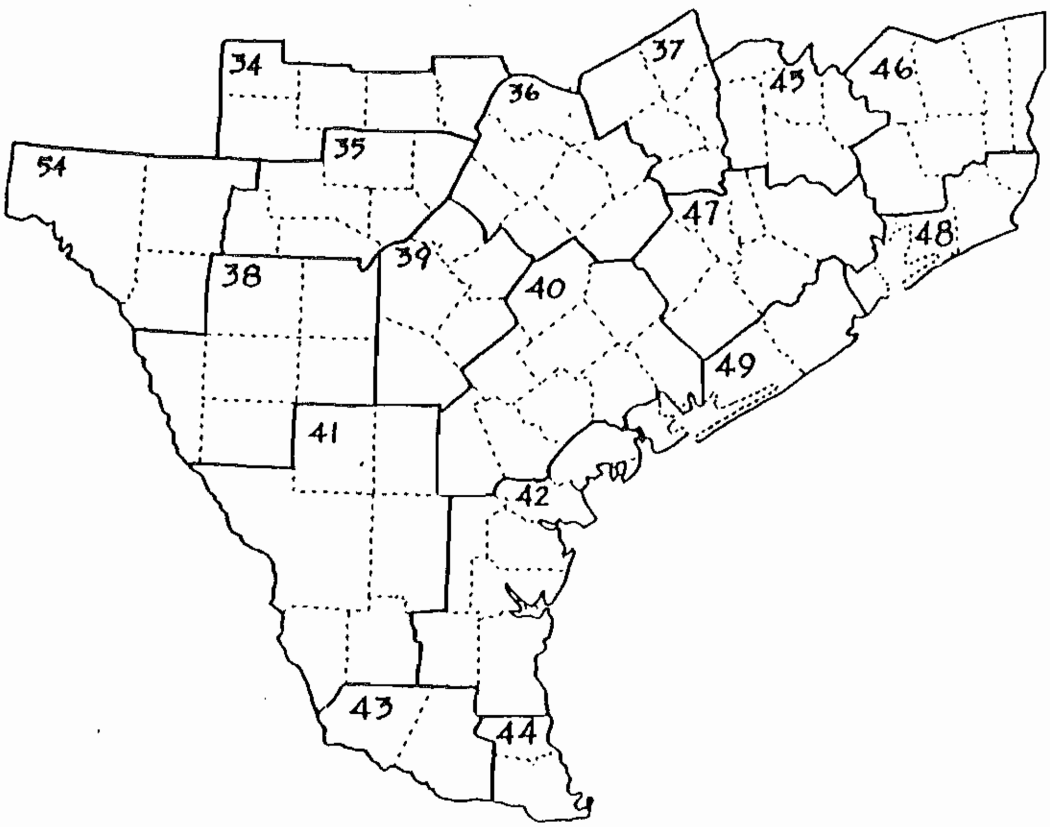
Suggested Index Threshold Values

For the probability of...

	GENERAL TSTMS			SEVERE TSTMS		
	LOW	MOD	HIGH	LOW	MOD	HIGH
Showalter	>2	+2 to 0	<0	≥2	-2 to -3	≤3
Lifted	≥1	-1 to -2	≤2	≥3	-3 to -4	≤4
K	>36	16 to 36	<16	N/A	N/A	N/A
Vert. Totals	<26	26 to 28	>28	N/A	N/A	N/A
Cross Totals	<18	18 to 20	>20	N/A	N/A	N/A
Totals Totals	<46	46 to 50	>50	<50	50 to 55	>55
SWEAT	<200	200 to 300	>300	<300	300 to 500	500 to 600

These are only suggested values that should be refined by local studies.

Appendix B
SOUTH TEXAS TORNADO CLIMATOLOGY
1950-1981



Percent Distribution of Tornado Characteristics by Zone

FORECAST ZONE	34	35	36	37	38	39	40	41	42	43	44	45	46	47	48	49	54
Max Wind (mph)	Percent of tornadoes																
below113 (F0,F1)	62	58	58	41	55	60	55	61	70	73	69	47	50	64	68	72	58
113-206 (F2,F3)	21	33	32	49	35	27	18	32	20	18	24	53	38	26	27	17	25
above 206 (F4, F5)	0	0	3	0	3	3	0	0	1	0	0	0	0	1	2	1	8
unknown	17	8	8	10	8	10	27	7	10	9	7	0	13	10	3	9	8
Path Length (miles)	Percent of tornadoes																
0 - 1 (P0)	34	39	44	37	43	60	43	61	63	36	55	26	63	62	62	59	42
1 - 10 (P1, P2)	34	25	30	24	40	21	24	25	23	36	28	53	23	24	30	25	33
over 10 (P3-P5)	10	6	9	20	3	4	1	4	2	9	3	5	2	4	1	4	8
unknown	21	31	17	18	15	15	32	11	12	18	14	16	13	10	7	12	17
Path Width (yards)	Percent of tornadoes																
below 500 (P0-P3)	62	50	66	67	63	79	57	75	83	64	66	68	75	82	89	81	75
500 - 1500 (P4)	17	17	11	12	18	1	6	7	4	9	3	11	6	4	2	4	0
over 1500 (P5)	0	6	4	2	5	6	2	4	0	0	3	0	4	3	1	0	8
unknown	21	28	19	18	15	13	35	14	12	27	28	21	15	11	9	14	17

SOUTH TEXAS TORNADO CLIMATOLOGY
1950-1981

FORECAST ZONE	34	35	36	37	38	39	40	41	42	43	44	45	46	47	48	49	54
	Number of tornadoes																
All occurrences	29	36	144	49	40	67	142	28	155	11	29	19	48	164	158	69	12
Killer tornadoes	0	0	3	1	2	1	0	0	1	0	0	1	0	6	6	1	0
Hour (LST)	Percent of tornadoes																
00 TIL 02	3	6	4	2	5	3	1	4	3	0	3	5	2	2	2	4	8
02 TIL 04	0	6	6	10	0	3	4	0	1	9	7	11	6	4	4	4	0
04 TIL 06	3	0	3	2	3	3	12	0	3	9	0	0	6	9	6	7	0
06 TIL 08	3	3	0	14	3	6	4	7	6	0	7	5	6	7	5	9	0
08 TIL 10	3	3	3	2	0	3	15	0	14	0	3	5	10	13	10	17	0
10 TIL 12	0	0	9	2	3	9	10	0	17	0	10	5	4	5	16	22	0
12 TIL 14	10	14	9	10	0	4	9	10	25	9	24	21	15	12	20	17	0
14 TIL 16	10	19	19	24	5	16	13	11	13	36	14	16	19	19	11	4	0
16 TIL 18	28	14	16	14	23	22	15	14	10	9	14	11	13	14	12	6	25
18 TIL 20	28	14	8	8	30	16	8	18	3	18	10	5	2	7	8	4	17
20 TIL 22	7	11	10	8	18	6	7	14	3	9	7	11	15	2	4	1	50
22 TIL 24	3	11	3	2	13	7	1	14	3	0	0	5	2	5	1	3	0
Month	Percent of tornadoes																
JAN	0	6	1	1	0	1	1	0	1	0	0	0	4	1	3	0	0
FEB	0	3	8	8	8	0	1	0	1	0	3	11	2	4	4	3	0
MAR	10	14	10	8	0	1	4	0	1	0	0	5	15	7	8	10	0
APR	0	3	22	27	15	10	8	7	5	18	10	5	13	9	6	9	25
MAY	41	58	20	33	43	31	25	39	16	0	28	16	17	15	10	13	67
JUN	7	3	4	6	3	3	6	11	15	0	24	5	4	13	15	12	0
JUL	3	0	3	2	3	6	4	0	8	9	10	16	6	6	15	4	0
AUG	14	8	11	0	18	10	11	11	19	36	21	5	6	7	10	10	8
SEP	21	3	11	2	3	19	31	18	22	27	3	5	13	18	11	33	0
OCT	3	0	5	2	5	6	7	11	8	9	0	21	17	10	11	3	0
NOV	0	3	3	6	5	7	1	4	3	0	0	0	0	5	4	3	0
DEC	0	0	1	4	0	3	1	0	0	0	0	11	4	4	2	0	0

Source of data is the NSSL log of severe storms. Note, however, that tornado statistics are affected by many factors. Since the mid-1970's, the training, organization, communications and areal coverage of spotter teams has been greatly improved. Hence, information since this time may be much more accurate than before the mid- 1970's. Population density also strongly affects these statistics. Tornadoes, especially small ones, are much more likely to be observed and reported when they occur near population centers. Note also that these summaries include tornadoes that were produced by both tropical events and non-tropical events.